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Edited by
O. M. B. BULMAN and S. R. NOCKOLDS

assisted by
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GEOLOGICAL MAGAZINE

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Age of the Wealden in North-Western Europe[1]

By P. ALLEN

ABSTRACT

The Wealden of the Weald is mainly post-Tithonian/pre-Aptian in age. The formation probably comprises the fullest non-marine record of the interval in north-western Europe, spanning most of Berriasian (Infra-Valanginian), Valanginian, Hauterivian and Barremian time. It is well suited to be the standard succession for purposes of direct correlation.

The term " Wealden " should, if used for beds outside the English basin, be applied only to those non-marine formations (1) clearly lying within the Berriasian–Barremian range and, if possible, (2) correlatable directly with the standard. Otherwise " Wealden facies " should be used.

Beds claimed as " Wealden " elsewhere in north-western Europe are examined in this light.

I. INTRODUCTION

OWING to its relict fauna and flora of Jurassic affinity the English Wealden has on occasions been assigned to the Jurassic System (see e.g. Strahan, 1898, p. 74). It is now, however, generally accepted as part of the Cretaceous. The Purbeck/Wealden boundary is thus implied to mark the transition from Jurassic to Cretaceous. In other words the boundary of Anderson's *setina* and *menevensis* zones (1939, 1940) may be approximately equivalent to the Upper Tithonian/Berriasian boundary of the marine Tethyan province. Concurrently, the idea has arisen that the Wealden of the Weald also bridges most of the time between the Upper Purbeck (Upper Tithonian ?) and Lower Greensand (Aptian).

How are these ideas affected by recent work ? Must they be revised in view of (1) Wolburg's report (1949*a*, 1950) that German " Wealden 4 " zone (pre-Lower Valanginian) ostracods occur in the top Weald Clay of Berwick, Sussex; and (2) Riedel's record (1941) of the Upper Tithonian or Berriasian ammonite *Blandfordiceras* in beds at Wahn (Germany) presumably overlying " Wealden " ? If these claims are correct, a substantial part of our Wealden must be Jurassic, with an enormous pre-Aptian time-gap at the top.

[1] Being an extended version of the Introduction to the *Sedgwick Prize Essay*, University of Cambridge, for 1952.

So far, and not excepting Grekoff's valuable review (1953), no attempt has yet been made to face up squarely to these and similar problems. This the present paper sets out to do.

II. Historical

The word " Weald " was used in naming the rocks in question at least as early as 1812 (Middleton's " Weald Measures "). But it was not until after the terms " Weald Clay " (Conybeare, *in* Conybeare and Phillips, 1822) and " Hastings-sand " (Fitton, 1824) had become common currency that Martin (1828, p. 9) proposed the Saxon-flavoured " Wealden " for the whole formation. By the 1870's this name was being generally used for similar beds in France and Germany (Lapparent, 1879; Struckmann, 1880).

The non-marine character of the beds, though long suspected (see, e.g. Conybeare, op. cit., pp. 120–1, 134–5; Mantell, 1822), was not fully recognized until 1824 (Fitton). By 1828 it was " generally believed " (Martin, p. 7).

The stratigraphical relations of the Hastings Sands and Weald Clay in the Weald were known to Conybeare in 1822 (op. cit.). He was also clear about the position of the Wealden between the Purbeck and Lower Greensand.

III. The Lower Boundary of the Wealden

(*a*) *Type Area*

The Purbecks of Sussex, at outcrop and in borings, have long been supposed to include representatives of the Lower, Middle and Upper divisions as defined in Dorset. Dr. F. W. Anderson of the Geological Survey tells me (*in litt.*, 1955) that recent work on the ostracods proves the point. The beds are conformable with the Wealden and the latter is therefore substantially post-Upper Purbeck in age.

The boundary of the Purbeck and Wealden in Sussex and Kent has always been chosen on grounds of mapping or logging convenience. Usually it is placed at a marked lithological change easily recognizable in borings or in the field. The macroflora and macrofauna, both freshwater or terrestial, do not change sufficiently for the purpose. Different criteria have been applied, the boundary being variously placed at (*a*) the " last [Purbeck] limestone "; (*b*) the " top of the bituminous (leathery) paper-shales with shell-seams "; (*c*) the " top of the green marls "; (*d*) the " lowest band of clay-ironstone nodules or sphaerosiderite at the base of the Fairlight Clays "; (*e*) " the base of the first [Wealden] arenaceous marls, silts, or sandstones "; (*f*) the " base of the green marls ", and (*g*) the " base of the mottled [Fairlight] clays ". Somewhat similar criteria were employed in Germany during the classic days of the nineteenth century, and the

" last limestone " was in use around Münder, Kapellenhagen and Duingen during 1948 by the German Geological Survey prior to receiving ostracod reports.

Anderson's work (1939, 1940) on the Purbeck/Wealden ostracod zones of the Henfield (Sussex) and other borings should eventually lend precision to fixing the field-boundary. The critical level is apparently the junction of his *Langtonia setina* and *Ulwellia menevensis* zones, but unfortunately the critical beds at the Durlston type-section in Dorset are not exposed (Anderson, *in litt.*, 1955). As yet the zones have not been applied to the lithological subdivisions at outcrop nor can the latter be recognized with certainty in any of the borings. But it is likely that the junction is low down in the Ashdown Series (Fairlight Clay facies), probably not very far from the Wealden base as traditionally defined. *U. menevensis* ranges at least up to the lower beds of the Wadhurst Clay (Allen, 1941, p. 371).

As long ago as 1886 Maillard discovered marine strata overlying Purbeck in the Cluse de Chaille (Savoie-Isère). The beds yielded ammonites recognized by H. Douvillé (Maillard, 1886, pp. 9–10) as *Perisphinctes* (*now Berriasella*) *lorioli* Zittel. Later, Joukowsky and Favre (1913) found a closely similar sequence at Mont Salève (Haute-Savoie). Kilian identified the same species of ammonite from marine strata immediately above the Purbeck there (p. 477). These records strongly suggest the contemporaneity of Purbeck and Upper Tithonian. Arkell (1955), who has examined the material, confirms the identifications of Kilian and H. Douvillé and fully establishes the correlation. The next step therefore turns on the precise age of the Purbecks in the Jura/Subalps-border area. So far, Upper Purbeck has not been rigorously proved, though it may be present (e.g. Carozzi, 1948), but both Middle and Lower Purbeck are known (Koert, 1898; Arkell, 1941 and 1955; Anderson, 1951; Grekoff, 1953). Hence our Upper Purbeck is certainly Upper Tithonian in age, and the bulk of the Wealden comes in as post-Tithonian, and truly Cretaceous. The non-marine *setina/menevensis* boundary therefore appears to correspond approximately with the Tithonian/Berriasian junction of the marine Tethyan province.

(*b*) *England Outside the Weald; Boulonnais*

Margins of the Basin.—Around the margins of the English basin the Wealden overlaps and oversteps the Purbeck and comes to lie unconformably on older rocks. Usually the basal beds are sandy and there are reasons for supposing that these normally represent part of the Hastings Beds. In the Boulonnais the junction with the Purbeck is strongly unconformable, thin Wealden cutting across strongly folded and faulted Lower Purbecks (*bononiensis* zone: Jones,

1880; Pruvost, 1925). Near Hardinghen it rests on Carboniferous and Devonian. This recalls northern Kent. The Boulonnais beds may therefore be Weald Clay in age, and in places they are indeed lithologically identical with the post-Hauterivian *argîles panacheés* of the Pays de Bray (see below). Unfortunately they have yielded no ostracods (Grekoff 1953), and can at present only be said to be post-Lower Purbeck in age.

Further along the basin margin the Wealden of east Kent similarly crosses the entire Jurassic succession, and comes to lie on Coal Measures at the northern rim of the coalfield. When the full evidence is assembled (e.g. Allen, 1954, p. 501, fig. 1), very considerable overlap within the Wealden is revealed. An upper argillaceous member, almost certainly Weald Clay, completely over-reaches the lower sandy (? Hastings) beds. Analagous relations are displayed near Oxford, where Wealden sands may lie on Lower Portland Beds (Arkell, 1947*a*, p. 131). Between Oxford and Dorset the Wealden outcrops at Swindon and in the Vale of Wardour are probably some way from the former margin of the basin. Those at Swindon comprise up to 3 feet of pebbly sand resting on Lower or Middle Purbeck (Arkell and Sylvester-Bradley, 1941). In Wardour the Wealden clays lie on Upper Purbeck.

Basin.—The arenaceous lower Wealden in the Isle of Purbeck and southern Dorset is conformable to the Upper Purbeck. It has so far yielded no ostracods, but presumably correlates, at least in part, with the Hastings Beds. To what extent however is doubtful, for the westerly attenuation of the overlying Wealden Shales (dying out west of Corfe) may be due either to erosion during the Aptian transgression or to lateral passage into a sandy facies. At present the balance of evidence favours the first explanation.

Nearer the centre of the basin, the scattered borings in Hampshire (Kingsclere, Portsdown—" K " and " P " in Text-fig. 1) and the Isle of Wight (Arreton) show conformable passages from Purbeck into sandier lower Wealden.

(*c*) *Continent of Europe*

North Germany, Holland.—Beds long correlated with the Wealden, and indeed so-called, occur in the Lower Saxon Basin (Text-fig. 1). Their classic locality in the " wald " country reminds one of the Saxon–Middle English derivation of " Weald " (wald, weeld). They usually follow conformably on the Serpulite, a formation varying from true serpulite to almost any lithological facies familiar in our own Purbeck. Recent work on the ostracods (Schmidt and Wolburg, 1949) has confirmed its suggested Middle Purbeck age (Koert, 1898; Arkell, 1941).

The German " Wealden " is thus clearly younger than part at least

of the Middle Purbeck, and probably includes near its base beds equivalent to our Upper Purbeck. During recent years ostracod zoning has been carried upwards through the entire succession (maximum thickness over 500 m.). Six major zones are recognized (Wicher, 1940; Wolburg, 1949*b*, 1950, and *in* Grekoff, 1953) and designated " Wealden 1 " to " Wealden 6 " upwards. Zones 1–5 are non-marine; No. 6 is brackish. " Wealden 1 " seems to be Middle Purbeck in age and " Wealden 2 " and " 3 " Upper Purbeck. The Purbeck/Wealden boundary thus probably lies between zones 3 and 4 or (Wolburg, 1950) in the upper part of 3. This is an interesting point, for it is about the level at which the ostracod-less Deister Sandstone and its equivalents come in around Hanover. On general grounds this may not be surprising, for the Dutch–German basin lies on the opposite flank of the same watershed (London–Brabant–Ardennes–Harz massif) as that in England. Hence, if topographic rejuvenation were the main reason for the arrival of sands and pebbles in both basins and if—as seems likely—Purbeck earth-movements were its cause, the base of the sands might well happen to be roughly the same age.

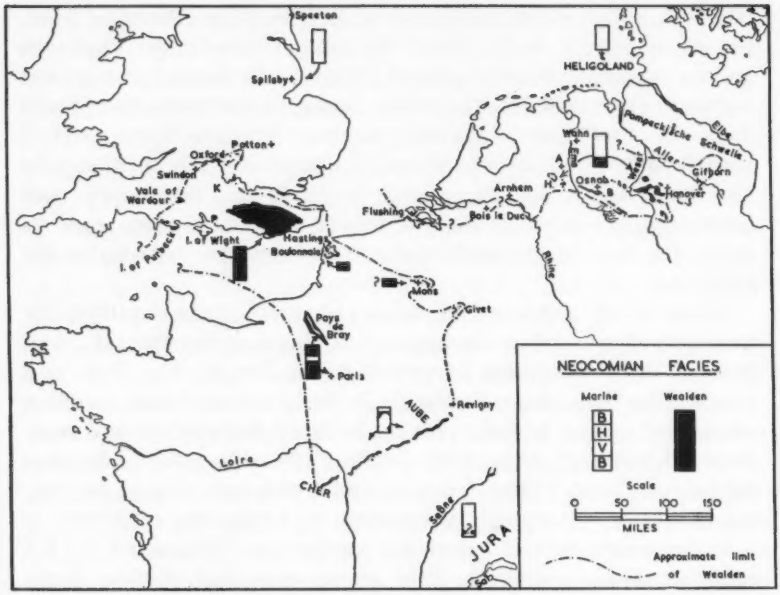
However, the sandstones around Osnabrück arrived earlier, for they yield abundant *Cypridea mammillata* Martin at Driemeier (1·5 km. N.E. of Tecklenberg) and Borgloh (" B " in Text-fig. 1). Their base seems to be " Wealden 1 " (Middle Purbeck) and they may therefore correspond, in part at least, with the ? Middle Purbeck silts and sandstones of Sussex (Topley, 1875; White, 1928). The latter lie between the " Greys " and " Blues ", not far above beds with *Ostrea*, *Mytilus*, etc. (Topley, p. 42) possibly representing the Cinder Bed of Dorset.

In the present state of knowledge the German " Wealden 1 "–" 3 " zones are clearly misnamed. They are no more than Wealden facies within the top Middle and Upper Purbeck, surprisingly like those of England, and should be labelled as such. Beds equivalent to our own are probably not present below the top of " Wealden 3 ". At or about this level the most voluminous sandstones and pebbly grits come in, a striking confirmation of the long-established correlation of the Deister and Hastings Sands (Struckmann, 1880; Wolburg, 1950).

It would, of course, be very satisfactory to be able to trace the non-marine beds laterally into fully marine Boreal strata and so back to the marine Neocomian of the east coast of England. A correlation independent of that with France and Switzerland (see below) might thus be effected. However, the two facies are not known to interlock anywhere, nor has any such locality been shown to exist so far as the writer is aware. (Despite reports of alternating marine and non-marine beds at Sehnde (Struckmann, 1880) and (in 1948) at Gifhorn (Text-

fig. 1).) Western and northern Holland appear to hold a key position in this respect. However, cores of certain West Dutch sediments attributed in part to the Wealden, and kindly supplied by the Nederlandse Aardolie Maatschappij, yield no ostracods and provide no further information on the subject.

As in England and the Boulonnais, the Dutch/German Wealden over-reaches the Purbeck along its margin and comes to lie on older rocks. In the frontier district west of the R. Ems (Schott, 1950, p. 247,



TEXT-FIG. 1.—Distribution of marine and Wealden facies of Neocomian in north-western Europe. *Borings*: K = Kingsclere, P = Portsdown, A = Adorf, H = Hengelo. Sal. = Mont Salève.

fig. 14) it oversteps the Jurassic completely and rests on the Muschelkalk near Hengelo (" H " in Text-fig. 1). In western Holland the Wealden is different in facies and indistinguishable from undated conformable sediments underneath, which lie on the Dogger. At a few places (e.g. Rossum, Ootemarsum, Otterstein, Deister) it is missing where perhaps Purbeck was present once. Along the northern margin of the basin similar relations have been proved by borings. Over the stable block north of the R. Aller (Pompeckj'sche Schwelle) and westwards towards the coast, the Wealden is thin and patchily distributed, lying on formations down to and including the Zechstein (Seitz, 1949, pl. iv).

Throughout the basin the position is greatly complicated by salt-domes. Wealden unconformities and pebble beds may result from local late Jurassic uprisings of evaporites, and cannot always be distinguished from those associated with normal marginal overstep. Local unconformities and pebble beds may thus occur well out in the basin (e.g. Wahn: Schott, 1950, p. 247). Nevertheless, subsidence clearly preceded the arrival of coarse detritus in the west. Transgression began at least by " Wealden 2 " times (Upper Purbeck), for the basal overstepping member of the Wealden has in some localities (e.g. Adorf: " A " in Text-fig. 1) definitely been assigned to that zone (Wolburg, 1949*b*). But these cannot be the movements that uplifted the massifs and speeded their rivers. For them we must look to the Osterwald earth-movements, proved by the conglomerates and unconformity at the base of the Serpulite.

It is interesting to compare these events in the areas so far considered:—

Date, in the main basin, of :	England	France (Boulonnais)	Germany	
			West (Adorf-Osnabrück)	East (*ca.* Hanover)
3. Initial transgression (subsidence)	Hastings ?	Weald Clay or early Hastings ?	Upper Purbeck	?
2. Arrival of coarse detritus	Early Hastings ; also (Sussex) late ? Middle Purbeck	Weald Clay or early Hastings ?	Late Middle Purbeck	" Wealden 3/4 " (= early Hastings ?)
1. Earth-movements (uplift)	?	Post-Lower Purbeck/ pre-Wealden	Post-Lower/ pre-Middle Purbeck	Post-Lower/ pre-Middle Purbeck

France.—Outside the Boulonnais the nature and date of the Wealden base are more doubtful and variable than in Germany.

In the Pays de Bray the thin Wealden (Lapparent, 1879; Dollfus and Fortin ; Thomas ; Lemoine) is clearly post-Portlandian *s.s.* (*senso anglico*). Whether it is also post-Purbeckian is unknown for no beds of this age have been proved and no ostracods found (Grekoff, 1953). The " Purbeck " pebble beds found at the base in the southern part of the area (Thomas) may well be Wealden. At the most it can only be said that the base is post-Lower Portlandian/pre-Hauterivian (see below).

Near Paris the old deep borings at Ivry and Pantin (Lemoine *et al.*, 1939) and the new ones at Beynes and Sezanne suggest Purbeck coming in beneath Wealden of Pays de Bray facies. Purbeck has been reported elsewhere in the Paris basin, but its precise age never recorded.

Along the south-eastern rim of the basin (Lemoine, 1911; Corroy,

1925), in the central outcrops (*ca.* Aube), the bulk of the attenuated Wealden is intra-Barremian. Its lower boundary sinks south-westwards, and beyond the Loire (Cher, Berry) probably reaches nearly to the base of the Barremian. North-eastwards, from Vassy to Revigny, the boundary also probably sinks. In addition, lacustrine clays and sands are developed sporadically lower down there, beneath the highly transgressive marine Hauterivian (Calcaire à Spatangues). Their base, lying on the Kimeridgian or Portlandian, is almost certainly Valanginian or later, and therefore rather younger than in the Weald.

Reports of Wealden from other parts of France remain unconfirmed. The most interesting are records of freshwater marls in the Berriasian of the Jura/Subalps border (Chambéry). These suggest a former extension of very early Wealden more like that of the Purbeck.

Belgium.—The lower limit of the " Wealden " in the Mons basin (Stevens and Marlière, 1944; Marlière, 1933, 1934, 1946) can at the most be placed as post-Coal Measures. Petrological work (unpublished) on the pebbles and other detritus has so far failed to prove the pre-existence of any later formations.

IV. The Upper Boundary of the Wealden

(*a*) *Type Area*

The Wealden of the Weald is demonstrably pre-Lower Aptian (*deshayesi* zone) [1] in age. Its upper boundary with the Lower Greensand is often gradational. Reasons are put forward later for believing that the junction does not represent a big time-gap. This suggests that the Wealden includes representatives of all the Cretaceous stages up to and including the Barremian. The contention is not finally proved but, pending further evidence, must be borne in mind when defining or correlating the Wealden.

(*b*) *England outside Weald; Boulonnais*

Around the margins of the English basin the contact is sharp and the overlying marine beds are normally younger than in the centre. Nevertheless, the pre-Aptian age of most of the Wealden underneath is almost certain. Along the southern flank of the London Platform the superincumbent beds seem usually to be Upper Aptian (Kirkaldy, 1939). In the Vale of Wardour and in Dorset west of Lulworth they are Middle Albian (Kirkaldy, 1939; Arkell, 1947*b*). Around Oxford and near Swindon the Wealden is preserved only as small outliers, presenting different problems in dating. However, it floors synclines of probable late Wealden or Lower Aptian age (Arkell, 1944, 1947*a*), and derived Wealden materials have been found in the Upper Aptian

[1] For the purpose of this paper the *recticostatus* zone is also included in the Aptian.

beds of Hartwell (Arkell, 1947*a*) and Potton (Keeping, 1883). The old idea that the Oxford Wealden is a non-marine facies of the Lower Greensand has now been disproved, for in the key area around the Miltons and Haseleys detailed mapping (Arkell, 1944, pp. 48–9, and pl. iv) has shown the Wealden and Lower Greensand to be separable by their lithology and stratigraphical and tectonic relations.

Nearer the centre of the basin, in the Isle of Wight, Hampshire and eastern Dorset, the boundary is again gradational (e.g. Worbarrow Bay), and the Wealden clearly pre-Aptian. In the Boulonnais it is also pre-Aptian and possibly slightly earlier than in England (Kirkaldy, 1937).

(*c*) *Continent of Europe*

Holland, Germany.—The Wealden regime of the Dutch/Lower Saxon Basin was destroyed by a much earlier marine inundation than that which eventually flooded England and the Boulonnais. Published information from Holland is scanty, but marine Valanginian fossils have been identified above " Wealden 6 " in the east of that country. In Germany the marine strata are very fully known and reported upon. The change is gradational in several places (e.g. north of Iburg, Hils district, etc.), brackish " Wealden 6 " being followed by "*Garneria*" (*Platylenticeras*)-shales with an abundant fauna. The age of the beds is Lower Valanginian, so the German " Wealden " lies almost entirely below the Valanginian. Its history was therefore over before Speeton times had even begun and also probably over before Hastings times had ended.

Lower Cretaceous geologists were therefore surprised when Riedel reported (1941) the Berriasian or Upper Tithonian ammonite *Blandfordiceras* in beds at Wahn 4 boring presumably overlying Wealden. However, Arkell (1955) considers that the ammonite fragments are almost certainly *Garantiana* derived from Middle Jurassic beds, and reports that consequent investigation of the cores by Wolburg failed to confirm the presence of any Wealden there.

France.—The uppermost Wealden (*argiles panachées* chiefly) in the Pays de Bray is overlain by Aptian, probably Lower (Lapparent, 1879; Dollfus and Fortin ; Thomas ; Lemoine). As a whole, therefore, the series (which includes a marine band) lies within the limits of the Sussex Wealden. This is also broadly true of the outcrops at the south-east of the Paris Basin. In the Aube country and north-eastwards to some distance beyond Vassy, the marine Vassy Red Bed delimits the underlying Wealden to within the Barremian. South-westwards (Berry, Cher) and north-eastwards (Meuse), the upper boundary may be later, for Aptian overlaps on to it in these directions (Lemoine, 1911; Corroy, 1925).

Belgium.—The upper boundary of the "Wealden" in the Mons basin is almost as unsatisfactory as the lower. The superincumbent strata are the marine Pommeroeul Beds at the base of the Meule, a Middle Albian formation (*dentatus* zone, Marlière, 1945). Only thin developments of "Wealden" are present and they could all equally well lie within the Aptian, or even within the early Albian.

V. Direct Correlation within the Wealden

(*a*) *England*

From the preceding remarks it is clear that the Wealden, properly defined, lies almost entirely in the Berriasian-Barremian range. But when one attempts to make direct correlations between different areas, serious difficulties arise. Not only is the palaeontological zonation incomplete outside Germany but much of the Wealden is in any case unfossiliferous.

Weald.—Within the English Weald the balance of evidence suggests that non-marine representatives of all the marine stages (Berriasian, Valanginian, Hauterivian, Barremian) may be present. This is supported as follows:—

(1) The Wealden/Aptian junction is conformable over a wide area from the Weald to Dorset and Hampshire.

(2) The contact is frequently gradational, there being no signs of marked erosion. (See also Kirkaldy, 1939, pp. 391–2, and (4) below.)

(3) There are no other indications of a lengthy break (such as beds rich in contemporary phosphate) at the junction. Indeed, the presence of derived Kimeridge pavlovids in both the top Wealden and basal Lower Greensand of the Isle of Wight (Casey, private communication, 1954) suggests just the opposite.

(4) Marine to brackish beds (*Ostrea*, *Mytilus*, etc.) occur constantly near the top of the Weald Clay in the Weald, Hampshire and Dorset. They may be said to anticipate the main Aptian transgression and thus to act as passage beds. (Barremian or *recticostatus*-zone ammonites have not yet been found.) This is supported by Casey's recent work (1955*a*, 1955*b*) on the main mass of the Weald Clay underneath, which shows a gradual transition downwards to the freshwater Hastings Beds via alternations of brackish strata (with *Filosina*) and freshwater beds (*Viviparus*, *Unio*, etc.).

(5) One of the brackish to marine lamellibranchs found near the top of the Wealden in the Weald (Arkell, 1947b, p. 176) and the Isle of Wight (Casey, private communication, 1955) is *Nemocardium* (*Pratulum*) *ibbetsoni* (Forbes), an Upper Barremian-Lower Aptian species.

(6) The Hastings Beds correlate well with German "Wealden 4" to "Wealden 6" (see below). The latter is pre-Valanginian in age. Hence

the 1,000 feet of higher beds in the Weald presumably represent the Valanginian to Barremian stages of the Lower Cretaceous.

(7) No really large gaps appear to exist *within* the thick and conformable succession of the Weald. Even the richest phosphatic bed (Top Lower Tunbridge Wells Pebble Bed: Milner, 1923; Kirkaldy, 1947) derived its nodules from Jurassic rocks (Allen, 1954).

(8) On petrological and other grounds (author's unpublished work; King, 1949) the Sussex Wealden was undoubtedly continuous with the Wealden of the Pays de Bray. The latter includes strata both older and younger than (Lower ?) Hauterivian (see below).

(9) The Wealden at the south-eastern rim of the Paris Basin includes beds of both Valanginian and Barremian age.

English Areas outside Weald, and Boulonnais.—Correlation between the scattered outcrops and borings in England outside the Weald and the Boulonnais awaits further work on their ostracods and megaspores. Meanwhile, it seems likely that the lower arenaceous strata in most places are equivalent to parts of the Hastings Beds and the overlying clays to the Weald Clay. Thus the fauna at Swindon is dominantly German (Arkell, 1948, p. 201). Cutting out of the upper shales westwards across Dorset may be due to sub-Aptian erosion, rather than to lateral passage of Weald Clay into a sandy marginal facies. The sandy outliers near Oxford, preserved in pre-Upper Aptian synclines, are also probably Hastings Beds for the same reason.

(*b*) *Germany, Holland*

Correlation of the East Dutch/German Wealden (zones 4–6 only) with the Hastings Beds, now suggested, rests on the following evidence:—

(*a*) The two formations develop their main sandstones at about the same level (Hastings Sands, Deister Sandstone and correlatives).

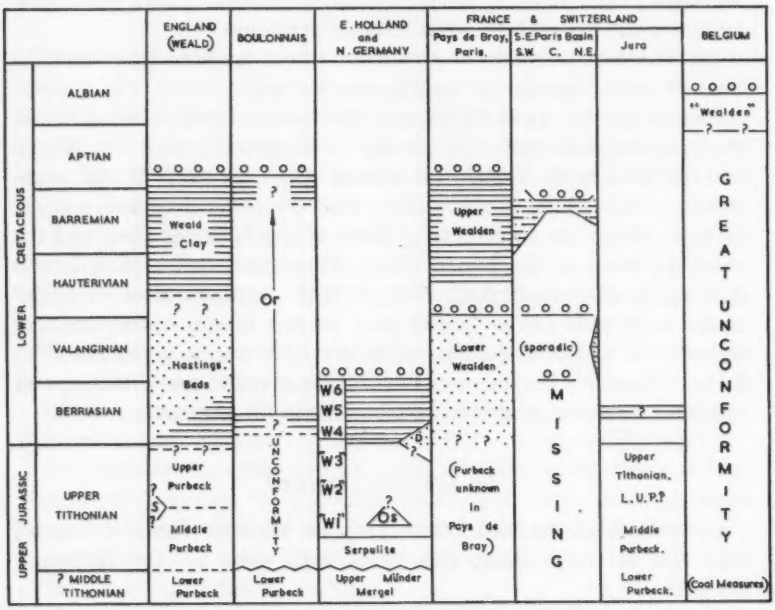
(*b*) Casey (1955*a*, 1955*b*) has recently obtained evidence suggesting a palaeontological break between the Hastings Beds and Weald Clay. The Hastings Beds contain abundant *Neomiodon*, a freshwater or freshbrackish water genus. This shell is so far unknown in the Weald Clay, the brackish strata of which contain *Filosina* Casey, a descendant of the Purbeck *Hemicorbicula* Casey. Casey recognizes *Neomiodon* as the characteristic shell of the writer's material from the "Wealden 5" of Germany (Georgsdorf 2 boring, depth 636 m.; Sedgwick Museum Nos. F 11957–11959).

(*c*) The basal Wealden preserved at Swindon is dominated by German species of lamellibranch (Arkell, 1948, p. 201).

(*d*) The remaining macrofauna and the macrofloras have long been known to be closely alike. Plant megaspores similar to those in the

Dutch Wealden (Dijkstra, 1951) have been obtained from the Hastings Beds of the Weald (Hughes, 1955).

The above correlation is, of course, quite different from that proposed by Wolburg (1949*a*, 1950), who recently examined an ostracod fauna from near the top of the Weald Clay at Berwick, Sussex. The sample contained species of "*Gomphocythere*" (*Metacypris*) and *Cypridea* closely allied to forms in the German "Wealden 4" zone. He therefore



TEXT-FIG. 2.—Proposed correlation of true Wealden (shaded) and of Wealden facies in north-western Europe. o o o o = marine transgression, D = Deister Sandstone and equivalents, Os = Osnabrück Sandstone, S = Purbeck sandstones of Sussex, C = central outcrops, L.U.P.? = lower Upper Purbeck ? Thicknesses not to scale.

concludes that the Wealden of the Weald may represent only a fraction of the German succession, itself entirely pre-Valanginian.

On this basis, it would follow that (1) an enormous time-gap (representing Valanginian, Hauterivian and Barremian) must exist at the top of the Weald Clay, and (2) a *substantial* part of the type-Wealden of Sussex must be Upper Purbeck in age! If no mistake has been made in the systematics, the ostracods concerned have failed and must either possess longer ranges than suspected or be otherwise unsuitable for long-range correlation. However, other forms, unknown

or rare in Germany, are present at Berwick. These, too, should perhaps be taken into consideration.

(*c*) *France*

Boulonnais.—The thin developments here (Pruvost, 1928) lack ostracods and cannot at present be correlated with the beds almost in sight across the Channel. On the one hand their structural position suggests Weald Clay; on the other, the lithology of the clays, while resembling the post-Hauterivian mottled clays in the Pays de Bray, also recalls Fairlight Clay.

Pays de Bray.—The Pays de Bray Wealden similarly lacks ostracods (Grekoff, 1953), but the succession is interrupted by a fossiliferous marine band (Lapparent, 1879 ; Dollfus and Fortin ; Thomas ; Lemoine). This provides a useful dating horizon, having for long been known to contain *Cardium subhillanum* Leymerie, *Panopea neocomiensis* d'Orbigny (considered by Woods to be a synonym of *P. gurgitis* (Brongniart)), *Astarte numismalis* d'Orbigny, *Nuculana scapha* (d'Orbigny), *N. mariae* (d'Orbigny), *Trigonia* sp., casts of gastropods, etc. Similar fossils in the Calcaire à Spatangues of the Aube country and Jura are associated with an undoubted Hauterivian fauna (cf. the " C " beds at Speeton). During 1949, *Corbula striatula* J. de C. Sowerby was obtained from chocolate-coloured clays exposed in a pit 1 km. south-west of Blacourt. This species has so far not been recorded from strata older than Hauterivian, and is therefore not inconsistent with a Hauterivian age. Dr. L. R. Cox of the British Museum kindly identified the specimens and searched the literature. The shells, together with other elements from the fauna at Blacourt, are now preserved in the Sedgwick Museum, Cambridge, under registered Nos. F 11950–11956.

In the Pays de Bray, therefore, the thin sequence is divided into a lower (arenaceous) division (= Hastings Sands ?) of pre-Hauterivian/post-Portlandian age, and an upper (argillaceous) group (= Weald Clay ?) of post-(? Lower) Hauterivian/pre-Aptian age. Presumably the lower division bridges Berriasian and Valanginian time. Indeed, it is sometimes referred to by French geologists as the " Valanginian-Wealden ". The upper division, mostly *argiles panachées*, is similarly often referred to as the " Barremian-Wealden ".

Twenty miles south-eastwards, the Paris borings (Lemoine *et al.*, 1939) suggest similar sequences. How far the tripartite succession continues beyond is unknown, for at the outcrops 100 miles on, the Aube Wealden is no more than a few thin sands and clays, *argiles bariolées* (= Weald Clay ?), mostly within the marine Barremian. Towards Revigny Wealden beds of Valanginian age (pre-Calcaire à Spatangues) occur, and south-westwards across Yonne to Berry,

almost the whole of the Barremian is gradually replaced by this facies.

The two transgressions and one regression of the Lower Cretaceous seas across France must certainly have affected the water levels in the Wealden swamps. At any one spot, perhaps as far north as the Weald, the movements might now be recorded as long-term changes in the non-marine facies of sedimentation. Could they, if recognized, be of use in correlation ? This is not impossible for south-eastwards, across the rim of the Paris Basin and in the Jura, the Hauterivian transgression and the Barremian regression are clearly reflected by deeper water and shallower water sediments respectively. In such a light the change from freshwater Hastings Sands to brackish Weald Clay may be of great significance.

The Wadhurst and Grinstead Clays may also reflect distant transgressions. But in this connection it must be remembered that the main Aptian transgression, at any rate, probably came up-Channel (Arkell, 1947*b*, p. 169) and not from across France. Hence the number of deeper water clays in the Weald may not necessarily accord with the number of transgressions in the Paris Basin.

Belgium.—As already pointed out, the Belgian Wealden facies may be later than anything known elsewhere. However, there is no clear evidence that the fauna and flora (Seward, 1900; Harris, 1953; Alvin, 1953) are essentially different from those in other parts of north-western Europe.

The rich and famous plant collections are nevertheless remarkable in the extraordinarily high proportions of drifted wood, chiefly cones and coniferous wood. Pieces of local waterside and fen-forest plants (ferns, *Weichselia*, etc.) are in a minority—the opposite of what is normally found in the Wealden of England, France and Germany. With the evidence of the delayed marine inundation, and the relations (Marlière, 1946) and very local petrology of the beds (author's unpublished work), this strongly suggests that the basin was isolated in some way. Probably it was a relatively small area of lake and fen partly enclosed within the forested slopes of the Brabant foothills.

VI. Summary of Conclusions

The tentative correlations drawn above are summarized in Text-fig. 2. They differ considerably from those advanced by Wolburg (1950) and Grekoff (1953). But no final solution can be hoped for until more is known about the rocks, particularly their molluscs, microfaunas and microfloras. The most promising lines of inquiry seem to be the lamellibranchs, ostracods and megaspores, and possibly also the otoliths of fish (Martin and Weiler, 1954).

VII. Acknowledgments

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DEPARTMENT OF GEOLOGY,
UNIVERSITY,
READING.

The Terms Teschenite and Crinanite

By J. F. G. Wilkinson

Abstract

Various aspects of the terms teschenite and crinanite are discussed, including their usage in relation to petrogenesis. It is suggested that the use of the term crinanite be discontinued.

Introduction

DURING the investigation of a differentiated teschenite sill near Gunnedah, New South Wales, a study of the literature revealed considerable variation in the meaning and significance of the terms teschenite and crinanite. "... there is little doubt that both names have been misapplied—crinanites have been recorded as teschenites and vice versa" (Hatch, Wells and Wells, 1949, p. 304). Basic intrusive rocks carrying analcite in varying amounts are not uncommon in the Tertiary igneous activity in Eastern Australia (Jensen, 1909; Jevons *et al*, 1911, 1912; Benson, 1911, 1913; Browne, 1924, 1927; Raggatt and Whitworth, 1932; Edwards, 1941, 1953; David, 1950, pp. 578–9). Apart from the uncertainty resulting from continued petrographic usage of the terms, the proposal of actual basic magma types characterized by each rock (Tyrrell, 1923, pp. 250–1; 1948, p. 196) carries with it important petrogenetic implications. Many of the difficulties of nomenclature have arisen from inconsistencies in the original definitions and their subsequent application. The following discussion embraces both a brief review of certain aspects of the terms, and suggestive remarks aimed at a reconciliation of the definitions and their petrogenetic applications.

Some Aspects of the Usage of the Terms Teschenite and Crinanite

The term teschenite was first used in 1861 by Hohenegger, who applied it in a very general sense to certain igneous rocks intrusive into Cretaceous strata in the neighbourhood of Paskau, Moravia. The teschenite of Tschermak (1869, p. 258) was a crystalline granular rock composed essentially of feldspar, hornblende, augite, and analcite. Both Teall (1888, p. 359) and Walker (1923, pp. 242–4) have summarized subsequent German and English variants of the term. Possibly the first recognition of teschenites in the Midland Valley of Scotland is due to Teall (1888, p. 191) in his description of Carcraig Rock, in the Firth of Forth.

The term crinanite was first applied by Flett (1911a, pp. 116–17) to certain rocks constituting a north-west system of dykes in Argyllshire,

of which Loch Crinan is the centre. In the same year, Flett (1911b, p. 42) gave a more concise definition of the type rock, namely, that " the crinanites, then, are dark-coloured fine-grained basic rocks consisting mainly of olivine, augite and plagioclase feldspar, with a considerable amount of analcite and zeolites." The earlier publication contained an analysis of the type (Table 2, Analysis II). Flett pointed out that the crinanites " present some analogies to the teschenites in mineral composition . . . from the latter they are distinguished by their finer grain, their perfect ophitic structure, the scarcity of hornblende and biotite and their occurrence as thin vertical parallel dykes " (teschenites characteristically occurring in the field as sills and laccoliths). However in his original description of the crinanites, Flett (1911b, p. 42) included rock types in which " the ophitic character is hardly visible." Walker (1934, p. 128) has pointed out that the use of the term " fine-grained " as descriptive of crinanites is unjustified as the coarseness of the grain shows a great deal of variation. The pegmatitic nature of certain of the Scottish crinanites has been commented on by Tyrrell (1923, p. 251). Structural relationships of crinanitic masses with the intruded sediments also show considerable variation.

Since the original statements regarding the nature of teschenites and crinanites, it has been recognized generally that the essential mineral constituents include basic plagioclase feldspar, titaniferous clinopyroxene, and analcite. However, subsequent definitions have included refinements in texture, the status of certain minerals as essential constituents, and so on. Some of the variations in definition are listed in Table 1. Except for Walker's review of the term crinanite in 1934, much of the available petrographic data on the rocks in question is unaccompanied by micrometric analyses, so that in many descriptions even the concept of certain minerals as " essential " is in itself rather vague. Whilst a division between both rock types could be based in certain cases on the status of the analcite as an essential or accessory mineral, this criterion would frequently be difficult of actual modal application, due to the presence in the rocks of a microcrystalline alkaline mesostasis, itself often highly analcitic, and the delicate analcitization of both plagioclase and alkali feldspars.

It is debatable whether the criterion of analcitization of the plagioclase feldspars is strictly correct in determining which rock type is developed (cf. Walker, 1923, p. 244; 1934, p. 124). Rocks may occur intimately within the one intrusion, in which both analcitized and unattacked feldspar may be present in different specimens. Such analcitization may arise if conditions in a certain part of the intrusion were favourable for concentration of the magmatic late-stage sodic solutions, in quantities sufficient for feldspar alteration to take place.

TABLE 1.—VARYING USAGE OF THE TERMS TESCHENITE AND CRINANITE

	Differences based on:—	Teschenite	Crinanite
(i)	Textural relations of the clino-pyroxene.	Coarse gabbroic through ophitic to fine granular (Johannsen, 1938, p. 226). Idiomorphic pyroxene (Holmes, 1920, p. 224). Pyroxene typically non-ophitic (Harker, 1895, p. 117).	"Perfect ophitic texture" (Flett, 1911b, p. 42) followed by Holmes (1920, p. 71).
(ii)	Presence of olivine as an essential constituent.	Olivine apparently not essential (Johannsen, 1938, p. 228; Holmes, 1920, p. 224). Olivine an essential constituent (Hatch and Wells, 1937, p. 226). Olivine is not constantly present (Hatch, Wells and Wells, 1949, p. 308).	Richer in olivine than teschenite (Tyrrell, 1913, p. 307). Abundance of *fresh* olivine (Eyles, Simpson and MacGregor, 1929, p. 378; Tyrrell, 1923, p. 252).
(iii)	Presence of barkevikite as an essential constituent.	Barkevikite essential (Holmes, 1920, p. 224; Iddings, 1913, p. 248; Johannsen, 1938, p. 228;). Barkevikite may or may not be present (Walker, 1923, p. 241; Hatch and Wells, 1937, p. 226).	
(iv)	Zeolite content.	Plagioclase feldspar analcitized (Walker, 1923, p. 244). Analcitization related to higher primary analcite content (Tyrrell, 1913, p. 307; Eyles, Simpson and MacGregor, 1929, p. 378).	"Secondary analcite . . . gradually replacing plagioclase" (Flett, 1911a, p. 117). Primary interstitial analcite but insufficient to cause analcitization of the plagioclase feldspar (Walker, 1923, p. 242). Poorer in analcite and alkali feldspar than teschenite (Tyrrell, 1913, p. 307).

Thus teschenitic types occur veining picrites and crinanites in the Garbh Eilean sill, Shiant Isles (Walker, 1930, p. 368).

Chemical differences between the two rocks are not particularly obvious (Table 2). Variation in Fe_2O_3, FeO and MgO in the average teschenite and crinanite is an expression of the modal and compositional variation of the olivine, clinopyroxene, and iron ore, and within the one intrusion may be brought about by simple fractionation crystallization. The presence of possible heteromorphism should not be over-

TABLE 2.—CHEMICAL ANALYSES OF TESCHENITES AND CRINANITES.

	I*	II	III	IV	V	VI	VII
SiO_2	41·42	43·94	46·99	44·69	45·57	44·8	45·4
TiO_2	3·14	2·45	2·20	0·46	2·41	3·0	2·6
Al_2O_3	15·07	14·03	15·92	14·17	14·95	15·3	16·1
Fe_2O_3	7·93	1·95	1·57	3·35	2·82	3·3	2·9
FeO	6·30	11·65	7·20	10·86	7·35	9·0	9·1
MnO	0·20	0·32	0·10	0·31	0·31	0·1	0·2
MgO	4·82	10·46	7·17	6·41	6·19	6·0	7·2
CaO	10·16	8·99	6·82	10·28	8·27	9·0	9·4
Na_2O	4·00	2·68	4·30	3·64	4·33	3·8	3·1
K_2O	1·98	0·33	1·38	2·01	2·16	1·9	0·4
H_2O^+	2·73	2·31	4·67	2·53	3·93	3·0	1·5
H_2O^-	0·27	0·85	0·69	1·05	0·97	0·2	1·4
Etc.	2·19	0·46	0·75	0·45	0·92	0·7	0·4
Total	100·21	100·42	99·76	100·21	100·18	100·1	99·7

Analysis I: Teschenite from Paskau, 25 km. west of Teschen, Mähren, Silesia. (Johannsen, 1938, p. 229.)

Analysis II: Crinanite, Slac nan Sgarbh, one mile north of Inver Cottage, Jura, Argyllshire. (Flett, 1911a, p. 118.)

Analysis III: Teschenite, Braefoot Promontory, Fife. (Campbell, Day, and Stenhouse, 1932, p. 354.)

Analysis IV: Crinanite, Howford Bridge, Mauchline. (Tyrrell, 1923, p. 254.)

Analysis V: Average of five analyses of the Late Palaeozoic teschenites of the Midland Valley of Scotland. (Tyrrell, 1923, p. 254.)

Analysis VI: Average of seven teschenites, Black Jack intrusion, Gunnedah, New South Wales.

Analysis VII: Average of six crinanites. (Walker, 1934, p. 126.)

* (No FeO quoted by Johannsen. FeO found by subtracting the oxides given from summation of 100·21.)

looked. The relatively high FeO and MgO contents of the type crinanite from Jura (Table 2, Analysis II) result from a higher modal olivine content (approximately 19 per cent by volume) than would appear to be average. In the past, emphasis has often been placed on the differences in the alkali and combined water contents of the two rocks, these constituents being higher in the teschenites. This would appear to be reflected in a higher zeolite and/or alkali feldspar content

in the latter rocks, but need not be necessarily so. Thus a higher soda content may result from an increase in the modal plagioclase feldspar in which normal zoning was more extreme than usual. Appreciable amphibole or mica would increase the combined water. Possibly potash may be significant, averaging 0·4 per cent in the crinanites and 2·2 per cent in the teschenites. In general, this would express itself mineralogically in a higher orthoclase or anorthoclase content in the latter rocks, so that with increasing K_2O, the teschenites grade into the glenmuirites of Johannsen (1938, p. 196). Depending, however, on the temperature of crystallization of the analcite, potassium may substitute for sodium in the analcite structure, both in intrusives, as in the Gunnedah teschenite sill, and in volcanic rocks (Larsen and Buie, 1938, p. 838). The potash content as revealed by a rock analysis consequently may constitute no indication as to the modal alkali feldspar present.

It is apparent that teschenite and crinanite may grade imperceptibly into one another. The critical basis of recognition of each type (particularly crinanite) is so fine that a particular rock may be designated as belonging to either type, according to the authority followed. The difficulties in nomenclature are well illustrated by the Dippin Sill, Arran, originally described as " olivine-analcime diabase " by Corstophine, and later by Harker (1903, pp. 112–14) as " teschenite ". Tyrrell (1913, p. 308) wrote that the rock was " more closely related to crinanite " but later (Tyrrell, 1928, pp. 118–19) recognized both teschenite and crinanite in the intrusion.

Recommendations and Conclusions

1. It is thought that retention of the term crinanite is fundamentally misleading inasmuch as the terms teschenite and crinanite as used by many writers are virtually synonymous. In his survey of those rocks which have been designated as crinanites, Walker (1934, p. 128) concluded that " it may be said that very few of the rocks described as crinanite bear any close chemical or mineralogical resemblance to the type example from Jura."

2. Retention of the term teschenite is favoured for the following reasons:

(*a*) It has priority of definition.

(*b*) In general its chemical and mineralogical affinities are such that it constitutes a well-recognized and widespread rock species.

(*c*) Uniform designation as teschenite of hypabyssal or plutonic basic rocks of alkaline character and carrying analcite as an essential constituent lends true emphasis to the frequent field association picrite-teschenite and focusses attention on an important petrogenetic

problem. The original picrite of Tschermak (1869) was an olivine-rich facies of a teschenite. The British Committee of Petrographical Nomenclature (1921, pp. 142–3) recommended that the term picrite be restricted to rocks related to teschenites, theralites, etc. The picrites and the teschenites are linked by the picroteschenite of Flett (1930, p. 67; 1932, p. 150) and Tyrrell (1948, p. 183).

3. Recognizing that textural variation was an unsatisfactory basis of classification of the rocks in question, Walker (1923, p. 244) proposed the following definition of teschenite on a mineralogical basis, covering most variations, namely that " teschenites are rocks consisting of plagioclase feldspar (which is partially or wholly analcitized), titanaugite and analcite. Barkevikite is often present, sometimes reaching a greater abundance than the titanaugite, and olivine is a frequent constituent." Accessory minerals may include nepheline, other zeolites, alkali feldspar, aegirine-augite, biotite, apatite and iron ores. The texture may be ophitic or otherwise. It is recommended that usage of this term be continued.

It is suggested that intrusive alkaline basic rocks with minor analcite, in amount comparable with the original Jura rock (2–3 per cent by volume), are adequately covered by the term " analcite-bearing olivine dolerites."

4. Since the broad concept of magma types was introduced by the authors of the Mull Memoir (Bailey *et al.*, 1924), it has been recognized that certain well-defined mineralogical and petrological trends may characterize the genetically related products of a particular magma type. Both teschenites and crinanites are members of the Mull Plateau magma type, or with subsequent broadening of the term, the olivine-basalt and alkali olivine-basalt magma types of Kennedy (1933) and Tilley (1950) respectively. There seems to be no fundamental petrogenetic reason for recognizing two separate parental magmas within the alkali series, namely the teschenitic and crinanitic (kylitic) magmas of Dr. Tyrrell (1923, p. 251; 1948, p. 196), if the basis of magmatic sub-division depends on differences in the degree of alteration of the olivine, the colour of the clinopyroxene, the amount of analcite and so on. Alkali olivine-basalt magma possesses those characteristics which entitle it to the status of a primary magma in its own right. Nockolds and Allen (1954, Table 17) have listed the compositions of parental magmas of the Scottish Tertiary, Mauritius, Hawaii, etc., well known areas of alkali-type activity. Whilst admitting to difficulty in the statement of a compositional definition of a magma type, the Mull authors (1924, p. 13) point out that it represents " an *average composition* around which actual rocks seem to group themselves." The origin of the varied members within the alkali group is probably to be sought in terms of crystal fractionation, contamina-

tion, etc. Problems dealing with the petrogenesis of certain alkali olivine-basalt differentiates will no doubt be clearer as more precise knowledge concerning the stability relations and factors governing the crystallization of analcite and nepheline becomes available (cf. Friedman, 1951, pp. 28–9).

Uniformity of nomenclature within alkali olivine-basalt magma is important in relation to the petrogenesis of certain of its differentiates:

(*a*) Differentiation trends of teschenitic and crinanitic magmas resulting in the production of felsic end-members are not as inflexible as originally proposed, namely to result in the formation of lugarite and analcite-syenite respectively. Walker (1926) has demonstrated the presence of contemporaneous segregation veins of analcite-syenite in the teschenite sill of Charlestown, Fife. In both the Gunnedah and other teschenites studied by the writer, a characteristic feature is the presence of a microcrystalline alkaline mesostasis, very similar in its composition to analcite-syenite. Such an intrusive association is analogous to the trachyte-olivine-basalt association of the oceanic islands.

(*b*) Davis and Hess (1949, pp. 857–8) have divided ultramafic rocks into three groups, based on earlier concepts suggested by Hess (1938). These groups are: (i) Ultramafic rocks of the basaltic magma suite; (ii) Ultramafic rocks of the secondary peridotite suite; (iii) Ultramafic rocks of the primary peridotite suite, typically located along the major axes of the island arcs and alpine mountain belts.

It is suggested that two groups be recognized within (i), whose petrogenesis should be critically considered in relation to their association with a particular magma type, namely:

(i*a*) Those ultramafic rocks located in the lower levels of sills and laccoliths of tholeiitic parentage, e.g. the Bushveld and Stillwater complexes. Orthopyroxenites and websterites may arise from a reaction relation between early-formed olivine and liquid.

(i*b*) The ultramafic picrites *sensu stricto*, found at various levels in teschenitic and related intrusions. Ortho-pyroxene is absent from these rocks. The olivine is frequently accompanied by a highly diopsidic titaniferous clinopyroxene whilst a basic plagioclase feldspar may also be present. Rocks of this type are usually richer in Al_2O_3, CaO, and the alkalies than members of the preceding group or (iii) above. The origin of such rocks in many cases probably results from processes more complex than the mechanism of simple gravitative accumulation of olivine crystals *in situ* (cf. Drever, 1952).

If the term picrite is to be used in a descriptive sense, e.g. as an indication of relative richness in olivine, an additional adjective indicative of its affinities with a particular magma type should be added, e.g. the term tholeiitic picrite-basalt.

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DEPARTMENT OF GEOLOGY,
THE UNIVERSITY OF NEW ENGLAND,
ARMIDALE, NEW SOUTH WALES.

Magmatic and Amygdaloidal Plagioclases

By M. J. Le Bas

(PLATES XIII AND XIV)

Abstract

Thermal metamorphism of zeolites in amygdales of basaltic rocks is shown to give glomero-porphyroblastic aggregates of plagioclase crystals. Plagioclase phenocrysts in basic rocks are shown to undergo but little change during similar degrees of metamorphism.

Introduction

DURING the investigation of the Carlingford Tertiary Complex in Ireland, it has been found necessary to examine the theories of genesis of porphyritic and glomero-porphyroblastic plagioclases in basic rocks. These have been the subject of much discussion (Bailey, 1924, p. 25, p. 155, p. 161; Bowen, 1928, p. 136; Harker, 1904, p. 50, p. 270, p. 329; Harker, 1951, pp. 103–5; Reynolds, 1951, pp. 101–2; Richey, 1930, pp. 86–8, pp. 288–9). The two more important and relevant processes are: large feldspars due to early precipitation from a magma (Bailey, 1924, p. 25; Bowen, 1928, p. 136; Richey, 1930, pp. 86–8), and feldspars derived by thermal metamorphism of lime-soda zeolites in amygdales (Bailey, 1924, p. 155, p. 161; Harker, 1904, p. 51, p. 329; Harker, 1951, pp. 103–5; Reynolds, 1951, pp. 101–2; Richey, 1930, pp. 288–9). This is not to say that they are mutually exclusive. The difficulty has been in knowing the distinctions between the final products of either of these processes. It is an important problem as it contributes data towards the solution of the magmatic–metasomatic controversy on the origin of many rocks.

An amygdaloidal origin for certain feldspars of the Slieve Gullion gabbros has been put forward by Miss Reynolds (1951). For supporting evidence she utilises facts shown in the Skye (Harker, 1904, p. 51) and Mull Tertiary Memoirs (Bailey, 1924, p. 155), where the transformation from zeolite to feldspar is demonstrated. This alone is insufficient to prove the metamorphic origin of these Slieve Gullion plagioclases. The solution to this problem is thought to have been found by a study of the textural relationships of the plagioclases involved and also that of their host rocks.

Petrology

The descriptions in this paper are based on observations from several hundred thin sections from Skye, Mull, Ardnamurchan, Slieve Gullion and Carlingford.

(a) *Zeolite to feldspar.*—The principal amygdaloidal minerals are

scolecite, prehnite, epidote and albite (M'Lintock, 1915, p. 15). The minerals calcite, hornblende, chabazite, thomsonite, heulandite and stilbite are also occasionally present in amygdales (M'Lintock, 1915, pp. 4–9). The zeolite forms the bulk of the amygdale and it invariably occurs in fibrous and often radiating masses. Should a feldspar be formed at the expense of such a sheaf-like aggregate of zeolite, it is possible that the resulting plagioclase would tend to follow this original texture.

The formation of these plagioclases is best studied in the Mull basalts. Here, lime-bearing plagioclase can be seen being formed in two ways: from the zeolite and from the albite. M'Lintock (1915, p. 21) noted that the albite is usually at, and normal to, the edge of the amygdale. The zeolite is in the centre, and prehnite with epidote are in between. As the grade of thermal metamorphism increases, so the growth of the plagioclase is stronger (Plate XIII, Fig. 1). No measurements have been made to determine the variations of compositions of the plagioclase with increase in metamorphism since any systematic results obtained would apply only to the particular rocks examined. The initial composition is the decisive factor and this may or may not be modified by metasomatism. The formation of the lime-soda plagioclase from zeolite may involve little more than the expulsion of water (Harker, 1904, p. 51).

A check on the intensity of metamorphism can be obtained from observing the state of the basalt itself. In general the plagioclase laths have only been clouded and slightly zoned at the margins, and the clinopyroxene altered in part to hornblende with the development of biotite in the proximity of the iron-ores (Bailey, 1924, p. 152). The plagioclase phenocrysts undergo a small change similar to that of the feldspars of the groundmass (Richey, 1930, p. 288). This indicates low grade thermal metamorphism.

(b) *The feldspar.*—In the next stage of thermal metamorphism, such as occurs near the contacts of gabbroic intrusions, the basalts have been variously described as pyroxene granulites, granulitic basalts, basic granular hornfels and " beerbachites ".

At Harta Corrie in Skye, a specimen of metamorphosed amygdaloidal basalt collected by Harker gives the connection between the granulitic basalts and the metamorphosed basalts as described by M'Lintock (1915). This specimen (No. 2639 of the Harker Collection, Cambridge) is a compact dark basalt with numerous white phenocrysts and amygdales averaging 3 mm. in diameter (Plate XIII, Fig. 3). In sections of this specimen, the matrix shows a ragged granular mass of stumpy plagioclases of composition about An_{55} (all compositions in this paper are from extinction angle determinations) with irregular granules of olivine, ortho- and clino-pyroxene each about 0·1 mm.

across. The pyroxene forms about 60 per cent and olivine less than 5 per cent of the bulk of the matrix. Scattered throughout the matrix are abundant crystals of iron-ore. There is also a little biotite and green hornblende.

The plagioclase phenocrysts in this specimen all show evidence of having undergone thermal metamorphism though they have retained their identity as single crystals of feldspar. They have been fractured. This is seen in the dislocations of the twin lamellae most of which have been subsequently annealed. Sometimes small grains of pyroxene are present along these lines of weakness (Wells, 1954, p. 387). Also as a result of the fracturing, granulitization may take place along a crack or at the end of a broken crystal (Harker, 1951, pp. 108–9). The edges of the phenocrysts show slight modification, though the original euhedral nature can still be recognized. The outline becomes crenellated, this being determined by the texture of the matrix. The zoning of the phenocrysts is very varied. It is usually simple, as any original oscillatory zoning tends to be destroyed with intense thermal metamorphism. That the oscillatory zoning should tend to disappear more readily than the normal zoning is a function of the width of the zones and the degree of difference in composition between the zones. In normal zoning these factors are larger and thus less likely to be eliminated. Due to the fracturing, the zoning is irregular but both normal and reversed are present. The composition of the zones varies from andesine to bytownite.

The plagioclases in the amygdales of this basalt from Harta Corrie have quite a different texture (Plate XIII, Fig. 2). These feldspars form fine-grained aggregates, the individual crystals of which may show elongation along *c*. The bulk of the amygdale is formed of an interlocking mosaic of plagioclase which is a normal metamorphic texture. At the centre of the amygdale the texture differs slightly. Here the plagioclase nearly always occurs in irregular laths with a strongly suggestive sheaf-like arrangement of the longer axes (Plate XIII, Fig. 2; Harker, 1904, plate xvii, No. 4). These laths measure, on average, $0{\cdot}75 \times 0{\cdot}4$ mm. The feldspar crystals in the surrounding mosaic measure 0·05 mm. in diameter. Again at the edge of the amygdales the feldspars tend to grow in crystals slightly larger than those of the mosaic and perpendicular to the wall of the amygdale. Whilst the former is thought to be following the original zeolite structure at the centre of the amygdale, the latter could be due to the original albite (M'Lintock, 1915, p. 15) occurring there.

These glomero-porphyroblastic aggregates are on the whole monomineralic with but very small amounts of iron-ore and ortho- and clino-pyroxene. Epidote and prehnite are rarely found at this grade of thermal metamorphism.

The composition of these feldspars is about An_{60} at their cores. The larger crystals in the centre of the amygdale show strong simple normal zoning. The smaller feldspars in the mosaic also show slight simple zoning. In the larger crystals, Carlsbad-albite type twins are well developed. The smaller crystals show less twinning and then usually of Carlsbad or albite type. Pericline twinning is not seen here.

With higher temperatures in thermal metamorphism the texture of the feldspars is further modified. The phenocrysts tend to become more homogeneous, though zoning, even oscillatory is still recognizable. Also prominent are the original euhedral characters, the fractures and the reticulate outlines.

In the amygdales, the feldspars increase their grain size. Instead of being represented by several hundred crystals of plagioclase as above, they are formed by only a hundred or less. These crystals homogenise by loss of zoning and the sheaf-like arrangement is destroyed. The result is a granulitic aggregate of nearly equidimensional feldspars giving a saccharoidal texture (Plate XIV, Fig. 4). The average grain size is $0{\cdot}5 \times 0{\cdot}3$ mm. The crystals show Carlsbad-albite type twins with very subordinate pericline type.

Good examples of this occur at Eilean Carrach in Ardnamurchan (Nos. 32705 and 59869 of the Harker Collection). These rocks appear to have been described as porphyritic gabbros (Richey, 1930, pp. 221-2) but the above specimens are undoubtedly of amygdaloidal origin as recently deduced by Wells (1954, p. 388). A basalt from Ardnamurchan (Richey, 1930, p. 289) also seems to have been misinterpreted. It is described as having a granular matrix with feldspar phenocrysts recrystallized as a mosaic. It is more probable that these mosaics were originally amygdales.

The increase in metamorphism described above is shown in the host rock which attains a hornfelsic texture. The minerals become cleaner, sharper in outline and coarser. Olivine now becomes an important constituent. It cannot be used in thermal metamorphism as an absolute determining factor of the grade because it is often an original mineral in the rock. But it is evident from the study of thin sections that the amount of olivine usually increases with higher temperatures. Whether this is due to a metasomatic effect besides a rise in temperature has not been determined.

Conclusions

It is shown that the end product of thermal metamorphism of zeolite amygdales is not the production of single large crystals of feldspar but the formation of an aggregate of individually homogeneous calcic plagioclase crystals. Original feldspar phenocrysts normally survive this metamorphism as single large crystals though

FIG. 1.

FIG. 2.

PLAGIOCLASE IN AMYGDALES.

they may show many signs of the thermal action and mechanical fracturing.

In the case of the White Gabbro forming Layer 5 at Slieve Gullion, the evidence given in this paper does not support the postulated origin from an amygdaloidal basalt (Reynolds, 1951, p. 101). The large feldspar crystals represent original plagioclase phenocrysts.

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DEPARTMENT OF MINERALOGY AND PETROLOGY,
CAMBRIDGE.

EXPLANATION OF PLATES XIII AND XIV

FIG. 1.—Amygdale in basalt lava. The dark portions (outlined in white) at the edge are the fine grained products of low grade thermal metamorphism of the basaltic matrix. Most of the amygdale consists of plagioclase, most coarsely grained at the centre and finest between the centre and edge. Some fibrous radiating prehnite occurs between the plagioclase laths at the centre. A few small crystals of epidote are also present. Crossed nicols. × 26. Beinn Fhada, Mull. (No. 72658.)

FIG. 2.—Amygdale (outlined in white) in high grade thermally metamorphosed basalt lava. At the centre are the large zoned and ragged plagioclase individuals showing slight sheaf-like structures. This is surrounded by a mosaic of small plagioclase crystals. Enclosing the amygdale is the granulitic basalt rich in iron-ore. Crossed nicols. × 33. Harta Corrie, Skye. (No. 2639.)

FIG. 3.—Plagioclase phenocryst and a fine grained glomero-porphyroblastic aggregate of plagioclase in a granulitic basalt. The phenocryst shows the crenellated outline, and the plagioclase aggregate (outlined in white) exhibits the sheaf-like development of the crystals. The host basalt shows plagioclase with clino- and orthopyroxene and iron ore. Crossed nicols. × 36. Harta Corrie, Skye. (No. 2639.)

FIG. 4.—Amygdale in very high grade thermally metamorphosed basalt. The amygdale is represented by a granular aggregate of plagioclase. The groundmass is coarser than in the above examples and olivine is abundant. Plane polarised light. × 35. Eilean Carrach, Ardnamurchan. (No. 32705.)

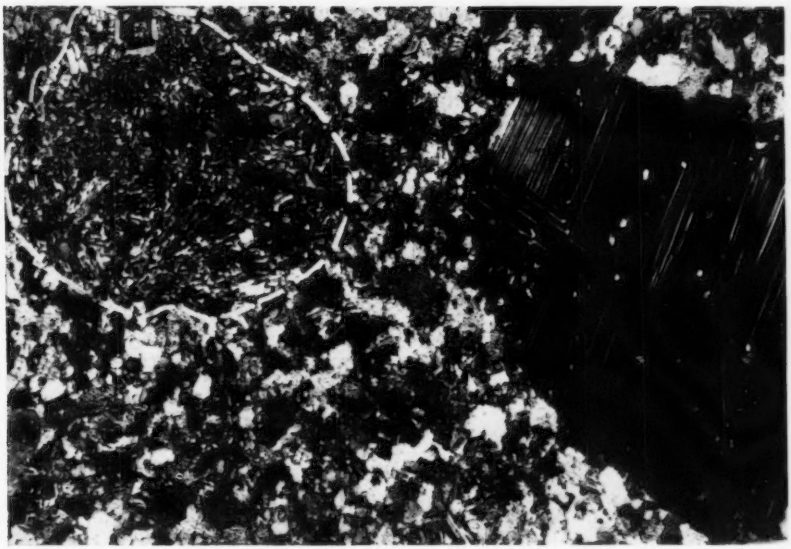


FIG. 3.

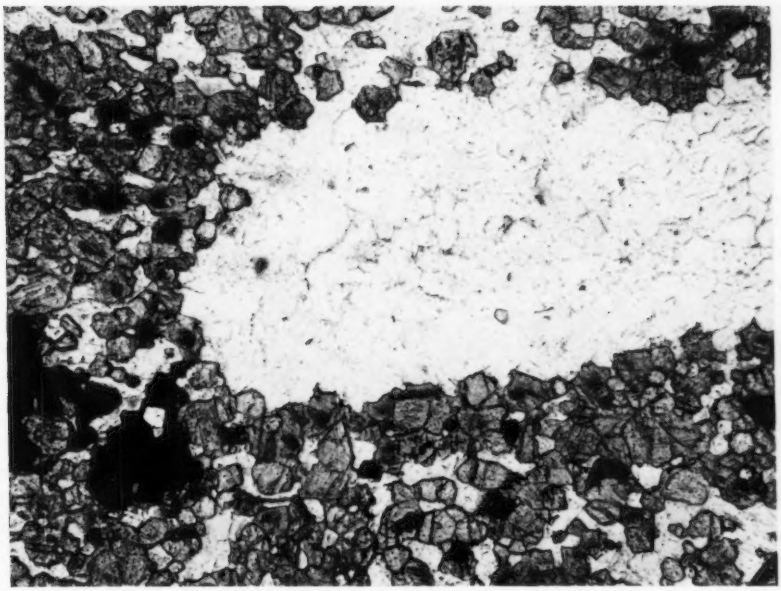


FIG. 4.

MAGMATIC AND AMYGDALOIDAL PLAGIOCLASE.

A Camptonitic Dyke Suite at Monar, Ross-shire and Inverness-shire

By JOHN G. RAMSAY

(PLATE XV)

ABSTRACT

A swarm of basaltic and camptonitic dykes in the Northern Highlands of Scotland has been studied. The dykes include monchiquite, olivine basalt, olivine dolerite, and camptonite, and a study of the relations of the various types in individual dykes suggests that the camptonites have arisen by the crystallization of late fractions from a magma of fairly normal dolerite type.

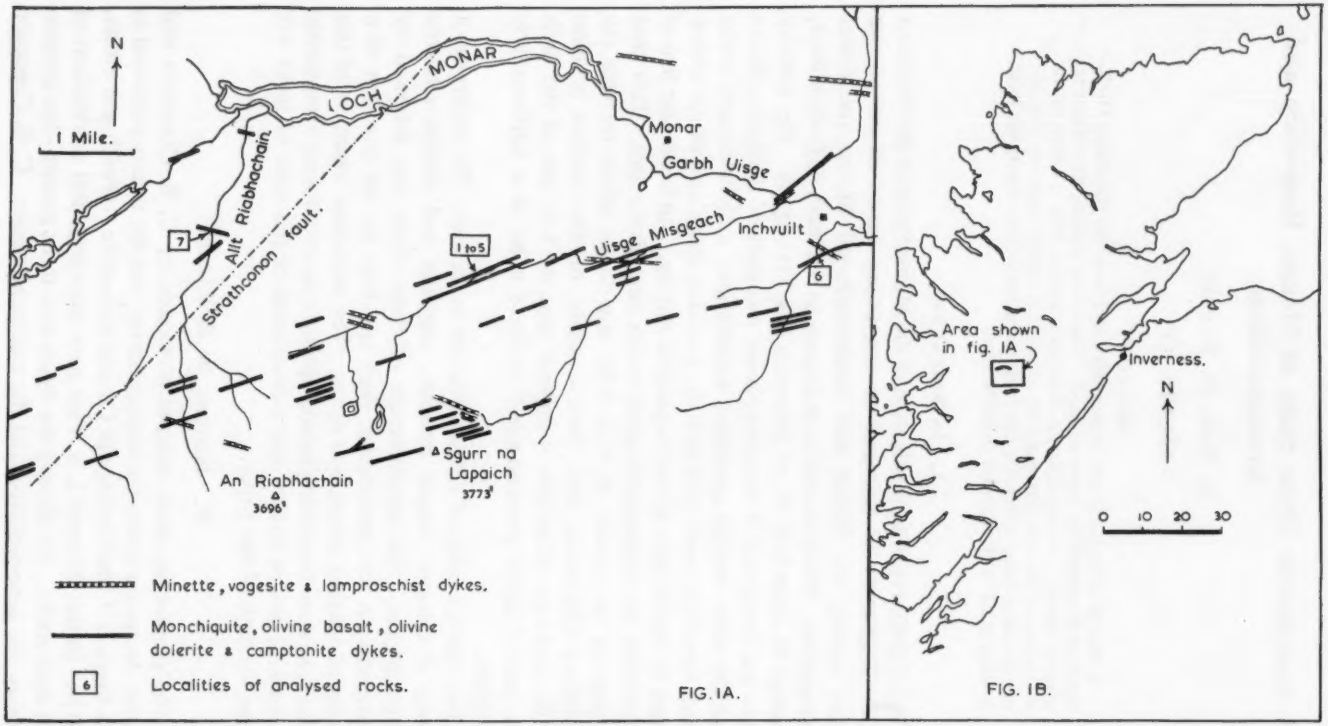
I. INTRODUCTION

DURING a recent investigation of the Moine Series in the Northern Highlands of Scotland an interesting group of camptonite dykes cutting the folded and metamorphosed Moine rocks was encountered. The dykes occur in an area to the south of Loch Monar, situated 30 miles W.S.W. of Inverness (see Text-fig. 1). The country rocks are composed of metamorphosed members of the Moine Series together with various gneissose, hornblendic, and calcareous rocks which have been correlated with the Lewisian rocks west of the Moine thrust by the Officers of the Geological Survey. An important line of dislocation, the Strathconon fault, passes through the area. This fault belongs to the system of N.E.–S.W. tear faults which traverse the Northern Highlands and from a study of the relative position of the rocks on either side it appears that the S.E. side of the fault has moved about two-and-a-half to three miles in a north-easterly direction.

Two distinct suites of dykes are found in the area. The earlier are a group of minettes, augite minettes, vogesites and various schistose lamprophyres. The emplacement of these dykes was followed by movement on the Strathconon fault and later by the intrusion of a group of basaltic camptonite dykes. The individual members of this dyke suite show interesting petrographical relations linking camptonitic rocks with olivine basalts and a description of this suite of rocks will form the bulk of this paper.

II. HISTORY OF RESEARCH

The ground has been mapped previously by C. B. Crampton and E. M. Anderson of the Geological Survey and the results published in the Memoir " The Geology of Central Ross-shire " (Peach and others, 1913). In this Memoir J. S. Flett gave petrographical descriptions of the dyke rocks. He divided the dykes into three groups—the lamproschists, the lamprophyres, and the camptonite suite. C. B. Crampton



TEXT.-FIG. 1.—Map of the Monar area showing location of dykes. Width of dykes not indicated.

found an olivine basalt of the camptonite suite which cut a minette dyke of the lamprophyre suite and Flett suggested that the camptonites were the later dykes, possibly of Tertiary age. Flett described the rock types within the camptonite suite and noted that " . . . very clear transitions between the different types occur in considerable numbers ". He did not put forward any petrogenic explanation for this fact. Since 1913 no further work has been carried out in the Monar area but work in other areas in the Northern Highlands has shown that a camptonite dyke suite is widespread (Harry, 1952, p. 139 ; Hartley and Leedal, 1951 ; Johnstone and Wright, 1951 ; Leedal, 1951 ; Read, 1931).

In 1939 Richey summarized much of the work which has been done on the dykes of Scotland. He compared the mica lamprophyre suite at Monar with the dykes associated with the " newer granites " found to the south of the Great Glen and he suggested that this suite was of Caledonian (Lower Old Red Sandstone) age. The petrographical similarity of the camptonitic dykes near Monar with the camptonite suite in the Midland Valley of Scotland led Richey to consider them part of that Permian suite.

III. The Mica Lamprophyre Dyke Suite

The dykes of this suite are earlier than and are cut by the camptonite dykes and appear unrelated to them. Only a brief description of these dyke rocks will be given below.

The mica lamprophyres vary in composition from normal minettes and vogesites to foliated green chloritic and hornblendic lamproschists. The dykes strike W.N.W. and often occur in groups. They are generally intruded into joint or fault planes and in places late movements on the joints have crushed and brecciated the dyke rocks. The average width of the dykes is 6 feet and a few reach 20 feet in thickness. A few dykes show dark contacts due to an abundance of chloritic material and occasionally they show spheroidal structures which weather out as nodules on exposed surfaces. These dykes are clearly earlier than the Strathconon fault for movements on parallel related faults have sheared and brecciated them.

In thin section corroded, bent, porphyritic crystals of biotite are common and have the pleochroic scheme X = pale buff, Y = Z = deep chestnut-brown, with darker outer zones. Idiomorphic porphyritic augites occur in the augite minette dykes. The augite is pale yellow-green in colour, faintly zoned with cores slightly darker in colour than the outer zones and with corroded crystal borders. They are broken and show irregular cracking. Idiomorphic crystals of apatite are common. These constituents are set in a confused matrix of brownish devitrified glass, altered plagioclase and orthoclase, with small flakes

of biotite and granules of augite, calcite, quartz, hematite, and magnetite. Occasionally the rocks are sheared to green biotite-hornblende schists. The petrography of these sheared lamprophyres at Monar has been described by Flett (in Peach and others, 1913, pp. 78–79) and very similar dykes at Ben Wyvis have received detailed description in the memoir accompanying sheet 93 of the Geological Survey of Scotland (Peach and others, 1912, pp. 120–126).

IV. The Camptonite Dyke Suite

(*a*) *Field occurrence*

Most of the dykes of this suite occur within a belt two-and-a-half miles wide which strikes W.S.W. through the valley of the Uisge Misgeach and continues on to the northern slopes of An Riabhachain (see Text-fig. 1A). This belt appears to be unaffected by the Strathconon fault since the dyke swarm is seen to continue across the fault plane without displacement. Individual dykes trend W.S.W. and commonly occur in parallel groups within the main swarm. Occasionally they trend W.N.W. and are sometimes intruded into the same fissures as the early mica lamprophyres. This W.N.W. trend is joint controlled, whereas the W.S.W. trend appears to be guided by the steep or vertical foliation planes of the Moine Series, the swarm being intruded into the vertical limb of a large synclinal fold. Individual dykes within the swarm cannot be followed for any great distance along the strike. The average width of the dykes is from 2 to 3 feet, a few reaching a thickness of 7 feet. The dykes frequently branch into two or more parallel dykes, the total thickness of the offshoots being greater than the thickness of the parent. The dykes sometimes show a perfect polygonal jointing.

In the field the rocks are generally fresh and dark grey or greenish-grey in colour. Porphyritic crystals of pyroxene and serpentinized olivine can be seen in hand specimens. Ocelli are sometimes visible as lighter patches, their centres filled with calcite, analcite, and phillipsite (often stained bright orange). The dyke contacts are always chilled to dark, fine-grained basaltic rocks without ocelli.

(*b*) *Petrography*

This suite is composed of monchiquite, olivine basalt, olivine dolerite, and camptonite. In thin section transitional forms are seen to be very common and intermediate types are more abundant than rocks strictly fulfilling the original definition of the named types described below. This monchiquite-camptonite suite is, in fact, largely composed of transitional forms and individual dykes frequently contain several varieties of rock.

(i) *Monchiquites*

These are fine-grained black rocks which in thin section characteristically contain idiomorphic phenocrysts of olivine and titanaugite up to 3 mm. in diameter set in a turbid, granular, isotropic base containing small idiomorphic violet-brown titanaugite crystals and magnetite dust and octahedra (see Plate XV A). The base is sometimes glassy but it is usually recognizable, in part at least, as clear anisotropic analcite. Occasionally small crystals of biotite are present in the matrix with X = pale buff, Y = Z = deep orange-brown. The rocks are characteristically without feldspar.

The corroded olivine phenocrysts are always completely pseudomorphed by serpentine, carbonates, and iron oxides. The pyroxene phenocrysts are zoned. The centres of the crystals are pale green in colour and show signs of magmatic corrosion before the acquisition of outer violet-brown titanaugite zones.

Ocellar structures are occasionally present containing long needle-shaped crystals of hornblende which has the pleochroic scheme X = pale yellow, Y = orange-brown and Z = deep chestnut-brown. Delicate skeletal magnetites are set in turbid analcite, with clear analcite and calcite filling the centre of the ocelli. Occasionally tiny andesine laths occur in the ocelli although absent from the surrounding rock.

(ii) *Olivine basalts*

In thin section these finely crystalline black rocks show phenocrysts of zoned augite (green cores and violet-brown titanaugite rims) and serpentinized olivine set in a dark glassy groundmass together with small titanaugite crystals, lath-shaped labradorite microlites (Ab_{30}), needles and octahedra of magnetite, calcite, and chlorite.

Rarely small ocellar patches are present containing deep brown hornblende needles and brown biotite. Zoned labradorite laths are common in these ocelli and are larger than those in the groundmass. They have cores of composition Ab_{30} and rims Ab_{40}. The ocelli have their centres filled with calcite and clear analcite. Skeletal magnetite is characteristic of the ocelli.

(iii) *Olivine dolerites*

None of the olivine dolerites shows a true ophitic texture and they have been separated from the olivine basalts on account of larger and more abundant plagioclase feldspar laths.

The phenocrysts are augite and serpentinized olivine. Plagioclase laths are larger and zoning is more prominent than in the olivine basalts. Olivine is the commonest porphyritic constituent. The borders of the crystals show signs of corrosion and the idiomorphic

crystals are always completely pseudomorphed by serpentine, chlorite, and occasionally chalcedony. The serpentine is accompanied by hematite and magnetite networks along the edge of the crystals and along irregular cracks. The augite phenocrysts are generally smaller than the olivine crystals. They are perfectly idiomorphic and are sometimes arranged in radiating fan-like groups. The augite is zoned and the violet-brown titanaugite crystals have a central core of pale green to pale straw-yellow augite. This central core commonly has inclusions of serpentinized olivine whilst the outer titanaugite zones have included magnetite dust and octahedra. Twinning is rare but hour-glass structures are common. The augite sometimes shows alteration to chlorite along cleavages and cracks.

The phenocrysts are set in an equi-granular matrix of small idiomorphic violet-brown titanaugites, plagioclase laths, magnetite, and occasionally small biotite flakes. The small pyroxenes have the optical characters of the outer zones of the porphyritic augites. Plagioclase laths are sometimes grouped in radiating fan-like aggregates. The cores are of composition Ab_{30} to Ab_{35} and the outer zones Ab_{60} to Ab_{70}. The zoning is perfectly continuous and the outer zones apparently show a more rapid compositional change than the inner zones. Magnetite is generally abundant as idiomorphic octahedra, as grains and as fine dust. Ilmenite grains show alteration to white powdery leucoxene. Interstitial translucent, glassy grey material is occasionally found ; sometimes it is recognizable as analcite.

Ocelli.—Most of the olivine dolerites are characterized by pale oval ocelli up to 5 mm. in diameter containing plagioclase laths, long narrow hornblende crystals and interstitial glassy material (see Plate XV C). Occasionally augite laths project from the groundmass into the ocelli but apart from this augite is never present. Hornblende is confined to the ocelli. The crystals are elongated, their length being up to ten times their width. The hornblende is a strongly pleochroic variety and is probably a titaniferous barkevikite, X = pale yellow, Y = orange-brown, Z = deep chestnut-brown, $Z > Y > X$. Its birefringence is low, $n_\alpha - n_\gamma = \cdot 008$; $Z \wedge c = 12°$; $2V_\gamma = 79°$ to $81°$. The plagioclase laths in the ocelli are always larger than those in the groundmass around the ocelli. They are continuously zoned, the composition of the cores is about Ab_{35}, whilst that of the rims is about Ab_{80}. The outer zones are always more sodic in composition than the outer zones of the plagioclase laths outside the ocelli. Magnetite dust occurs in the ocelli but magnetite is usually present as long feathery, skeletal and dendritic crystals frequently included within the outer shells of the plagioclase. The centres of the ocelli are filled with calcite and clear anisotropic analcite generally surrounded by a rim of chlorite. These ocelli appear to have been derived from the late-stage liquid residuum

segregated from a crystallized magma. Crystallization has gone on from the margin of the ocellus inwards leaving a miarolitic cavity at the centre which became filled with calcite and analcite.

The ocelli become more abundant and their borders become vague as the ocellar olivine dolerites pass into camptonites.

(iv) *Camptonites*

The camptonites have nearly all the mineralogical characters of the ocelli just described and this rock type (see Plate XV B) appears to be derived from the late stage liquid residuum after the crystallization of dolerite.

Porphyritic serpentinized olivine and green augite with violet-brown titanaugite outer zones are identical with the porphyritic constituents of the previous rock types. Porphyritic dark brown hornblende is common and the crystals frequently have a central core of titanaugite. This hornblende has all the optical properties of the hornblende found in the ocelli of the olivine dolerites. Small porphyritic crystals of violet-brown titanaugite are present and frequently show a peripheral growth of dark brown hornblende. Plagioclase laths are large and are continuously zoned with cores of composition Ab_{35} and rims Ab_{80} to Ab_{90}. Only in one dyke has a discontinuity in zoning been seen where a slight absorption of the early formed plagioclase has taken place. Magnetite is abundant as large feathery crystals and is also found in the form of dust and octahedra.

(*c*) *Variations in rock type in individual dykes*

The smaller dykes are the more uniform in composition and consist of the rock types monchiquite and olivine basalt. The larger dykes are very variable in composition. The rock types at the contacts of the larger dykes are generally feldspar-free or feldspar-poor varieties—monchiquite and olivine basalt—whilst the rock at the centre is usually ocellar olivine dolerite, doleritic camptonite or camptonite.

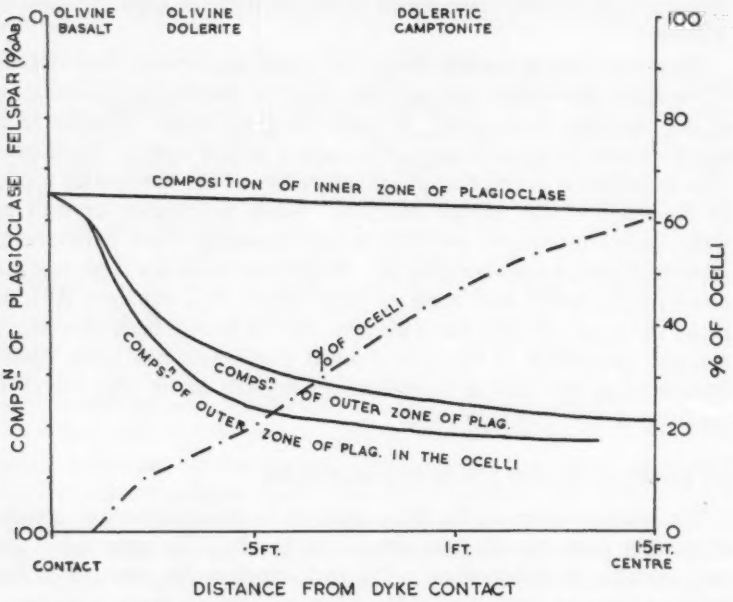
Xenoliths are rare in these dykes but one dyke in the Allt Riabhachain shows small xenoliths of dark olivine basalt in a camptonitic dolerite.

The mineralogical variations in one dyke 3 feet in thickness in the stream section of the Uisge Misgeach three-and-a-half miles above its junction with the Garbh Uisge have been studied in some detail. In the field this dyke has black chilled basaltic edges showing slight banding and schlieren near the contact and a dark grey central portion which is more coarsely crystalline.

The compositional changes in the plagioclase feldspar of the Uisge Misgeach dyke are illustrated graphically in Text-fig. 2. The plagio-

clase feldspar at the edge of the dyke is present as microlites of composition Ab_{37}. These increase in size towards the centre of the dyke and become increasingly zoned. The zoning is continuous and on rotation between crossed nicols the zoned portion shows continuous change in position of extinction. The plagioclase acquires outer shells which are progressively more sodic towards the dyke centre whilst the cores of the crystals remain fairly constant in composition (Ab_{35}).

Small ocelli appear one-and-a-half inches from the dyke contact. The plagioclase in these small ocelli is larger, more abundant and



TEXT.-FIG. 2.—The compositional changes in plagioclase feldspar found in the Uisge Misgeach dyke.

richer in Ab content than the plagioclase in the groundmass of the rock around the ocelli. The ocelli increase in size and abundance towards the centre of the dyke and eventually make up 60 per cent of the total rock volume. Six inches from the dyke contact the plagioclase laths in the ocelli are almost double the size of the plagioclase laths outside the ocelli. The outer shells of these plagioclase laths have a composition Ab_{77} in contrast with the outer shells of the plagioclase outside the ocelli with composition Ab_{68}.

Serpentinized olivine and small porphyritic crystals of green augite persist practically unchanged throughout the dyke and these two constituents appear to have existed as crystals in the original magma.

The augite acquires outer violet-brown titaniferous zones of increasing thickness towards the dyke centre. The size of the idiomorphic titanaugite crystals in the groundmass also increases towards the centre of the dyke.

Hornblende is absent at the dyke contact. It appears as small needle-shaped crystals in the small ocelli one-and-a-half inches from

TABLE 1

	1	2	3	4	5	6	7	V	VII
SiO_2 .	45·4	44·6	45·7	43·5	43·8	43·1	41·3	45·24	46·46
TiO_2 .	2·2	2·3	2·1	2·2	2·3	1·9	3·0	3·42	4·01
Al_2O_3 .	14·2	14·4	15·0	13·7	13·7	14·8	13·7	14·65	14·45
Fe_2O_3 .	4·1	4·7	3·4	5·2	4·9	4·1	7·7	3·38	3·79
FeO .	6·2	6·2	6·4	5·8	6·6	4·7	6·7	10·93	9·87
MnO .	0·2	0·2	0·2	0·3	0·2	0·2	0·2	0·17	0·15
MgO .	9·8	8·1	8·2	11·2	8·5	7·0	5·9	5·18	4·59
CaO .	7·4	10·1	9·7	10·2	9·5	14·7	11·9	8·98	7·48
Na_2O .	2·3	2·5	3·2	2·0	3·2	2·4	2·8	2·97	4·23
K_2O .	1·5	1·7	1·7	1·7	2·1	1·4	1·8	1·69	1·96
P_2O_5 .	0·6	0·6	0·7	0·5	0·7	0·7	0·7	0·65	0·82
H_2O .	4·5	3·7	3·8	4·3	3·3	3·6	3·5	2·67	2·05
CO_2 .								0·23	trace
BaO .								0·06	
S .								0·16	
Total .	98·4	99·1	100·1	100·6	98·8	98·6	99·2	100·32	99·86
Distance from contact in inches	0	2	6	12	18				

1 to 5: A series of rocks from a dyke in the Uisge Misgeach, 3½ miles above its junction with the Garbh Uisge. Analyst: R. C. Sinha.
6: Olivine basalt dyke, half mile S.W. of Inchvuilt. Analyst: R. C. Sinha.
7: Camptonite dyke, in the Allt Riabhachain, 1½ miles south of Loch Monar. Analyst: R. C. Sinha.
V and VII: Doleritic camptonite and camptonite from Skaergaard peninsula (Vincent, 1953, Table III).

the dyke contact. It increases in abundance and in size with the ocelli and reaches a maximum development at the dyke centre.

Chemical analyses of a series of rocks from this dyke are given in Table 1. The author would like to express his thanks to Dr. R. C. Sinha for the chemical analyses of the dyke rocks. The analyses show that the magma was a fairly normal basaltic type rich in water and potash. The only significant chemical changes in the rocks from the dyke edge (an olivine basalt) to the dyke centre (a camptonitic dolerite) are an increase in the proportion of soda and a noticeable increase in the percentage of potash. When these analyses are compared with

analyses of dolerite and camptonite dykes from Skaergaard (East Greenland) given by Vincent (1953) it is seen that there is a close similarity. The narrow chemical variation in the Monar dykes is a feature of the Skaergaard dykes and Vincent's remark, ". . . the changes are in the mineralogy of these rocks rather than in their chemistry " could equally well apply to the Monar camptonite suite.

(*d*) *Petrogenesis and Conclusions*

A petrographical study of the various rock types in individual dykes suggests that the camptonites have arisen by crystallization of late fractions from a magma of basaltic type. The continuous variation of rock type in the Monar dykes from monchiquite and olivine basalt at the contact to ocellar camptonitic dolerite and camptonite at the dyke centre forms a genetic link between these rock types.

In the smaller dykes the magma was intruded into the dyke fissure and was chilled before the crystallization of plagioclase feldspar could take place. The resulting monchiquites have a glassy base, often recognizable as analcite. The chilled selvage rocks always contain phenocrysts of serpentinized olivine and altered green augite and it would appear that these were porphyritic constituents in the magma.

In the larger dykes the greater volume of hot magma appears to have so heated the wall rocks that rapid chilling of the outer part of the dyke did not occur and small labradorite laths or microlites crystallized from the magma at the edge of the dyke. The magma solidified from the contact inwards and the crystalline constituents increase in size towards the centre. The plagioclase laths are continuously zoned and the late crystallizing liquids have deposited albite-rich outer shells which become more sodic towards the centre of the dyke. This is the normal product of crystallization from an undisturbed magma.

The interstitial liquids frequently segregated into ocelli, leucocratic patches of lamprophyric camptonite rock surrounded by more normal doleritic rock. The plagioclase laths in the ocelli are larger and have outer shells richer in soda than the laths outside the ocelli. It would appear that the plagioclases in the ocelli acquired sodic outer shells when those outside the ocelli had ceased to grow. The late magmatic fractions were concentrated towards the centres of the dykes. The camptonitic ocellar patches increase in size and abundance towards the centre and in the thickest dykes the central portion is camptonitic in character and similar to that of coarsely crystalline ocelli. The camptonitic segregations appear to have crystallized from late-stage interstitial liquids in a partially crystallized doleritic magma. In the thickest dykes gradual solidification of the magma from the edge towards the centre concentrated the late-stage liquids into the central

TABLE 2.

Rock Type	Pyroxene		Olivine	Plagioclase		Hornblende	Magnetite	
	a. Green augite.	*b.* Violet-brown Ti-augite.		*a.* In ground-mass.	*b.* In ocelli.		*a.* In ground-mass.	*b.* In ocelli.
Camptonite	Green porphyritic crystals remain approximately constant in size throughout the rock types. Decreasing in abundance. →	Outer shells on green augite. Large idiomorphic crystals.	Corroded serpentinized crystals decreasing in abundance. →	Large laths. Cores Ab_{35} to $_{40}$. Rims Ab_{80} to $_{90}$.		Elongated robust crystals commonly around Ti-augite.	Large feathery crystals. Dust and octahedra.	
Olivine Dolerite		Outer shells on green augite. Idiomorphic crystals.		Laths. Cores Ab_{35}. Rims Ab_{60-70}.	Large laths. Cores Ab_{35}. Rims Ab_{80}.	Long needles in ocelli only. Sometimes around Ti-augite.	Octahedra and dust.	Large feathery and skeletal crystals. Dust.
Olivine Basalt		Outer shells on green augite. Small idiomorphic crystals.		Microlites. Ab_{30}.	Small laths. Cores Ab_{30}. Rims richer in Ab content.	Small needles in ocelli only.	Octahedra and dust.	Feathery and skeletal crystals. Dust.
Monchiquite		Outer shells on green augite. Small idiomorphic crystals.		None.	Rare microlites.	Small needles in ocelli only.	Small octahedra and dust.	Feathery and skeletal crystals. Dust.
Magma (Basaltic Type).		None.		None.	None.	None.	None ?	

portion of the dyke where they eventually crystallized as camptonitic rock.

A relationship between camptonite and basaltic rocks has been noticed by other workers. E. B. Bailey discusses camptonite dykes in the Mull Memoir and remarks " . . . when a large collection of slides is examined, it becomes evident that the camptonites are petrologically united with basalts of a type such as undoubtedly figures largely among Tertiary lavas and intrusions of Mull ". (Bailey and others, 1924, p. 378.) In this same memoir H. H. Thomas describes the petrography of the camptonite dykes and concludes : " Certain of the dykes have, in the main, distinct doleritic affinities, but present an ocellar structure of which the ocelli are lamprophyric both in composition and in the arrangement of the constituent minerals." (Bailey and others, 1924, p. 382.) Vincent (1953), in a study of the relationship of various camptonite and dolerite dyke rocks in East Greenland, has also brought forward evidence based on mineral textures and a dyke intrusion sequence that suggests that the camptonitic rocks have originated from late fractions of a parent basic magma of normal doleritic composition.

The mineralogical characters of the rock types are summarized in Table 2. This table places the various types in a supposed petrogenetical order from the magma through the finely crystalline monchiquites and olivine basalts to the late magmatic crystallizations of camptonitic rock.

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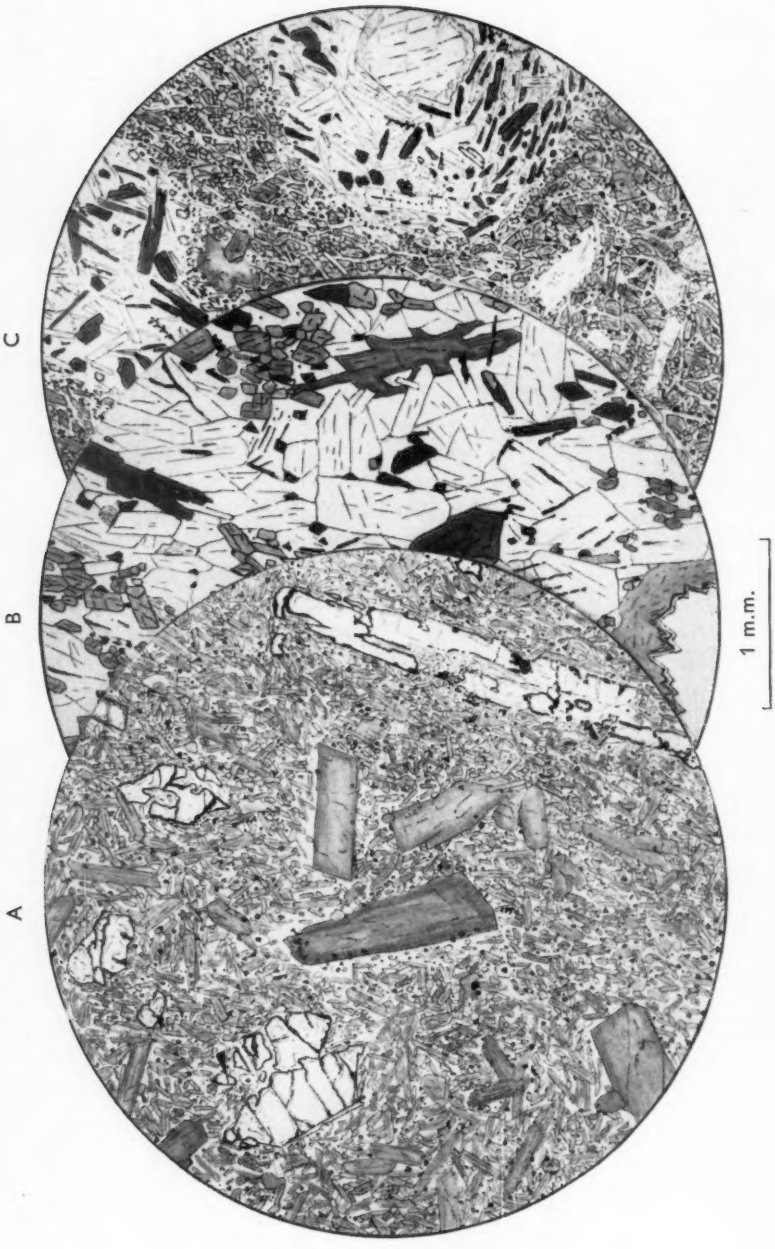
DEPARTMENT OF GEOLOGY,
IMPERIAL COLLEGE,
LONDON, S.W. 7.



DYKE-ROCKS AT MONAR.

EXPLANATION OF PLATE XV

A. Monchiquite, two-and-a-half miles N.W. of An Riabhachain. Idiomorphic phenocrysts of serpentinized olivine and zoned titan-augite are set in a glassy analcitic base together with small titan-augites, magnetite, and a little biotite.

B. A camptonite from the central portion of a dyke in the Allt Riabhachain. This rock has been chemically analysed (Table 1, No. 7). The rock consists of porphyritic serpentinized olivine, titan-augite and dark brown hornblende with large laths and plates of zoned plagioclase feldspar and magnetite.

C. An ocellar camptonitic dolerite from the Garbh Uisge 200 yards above its junction with the Uisge Misgeach. An equigranular matrix of small augites and labradorite laths with phenocrysts of olivine and titan-augite. Pale oval ocelli contain long narrow hornblende crystals, zoned plagioclase laths and dendritic magnetites and have a central portion made up of calcite and analcite.

The Mode of Occurrence of Sillimanite in the Glen District, Co. Donegal

By C. F. Tozer

Abstract

Sillimanite produced by the contact action of the Donegal Granite shows varied textural and time relationships with the other minerals of the schists in which it occurs. Early sillimanite was formed in the main stage of metamorphism by destruction of biotite, K, Fe, and Mg being expelled in the process. A later influx of new K from the granite led to the growth of muscovite metacrysts. This was associated with destruction of part of the early sillimanite, recrystallization of part, growth of a new crop of sillimanite needles, and formation of muscovite-sillimanite intergrowths.

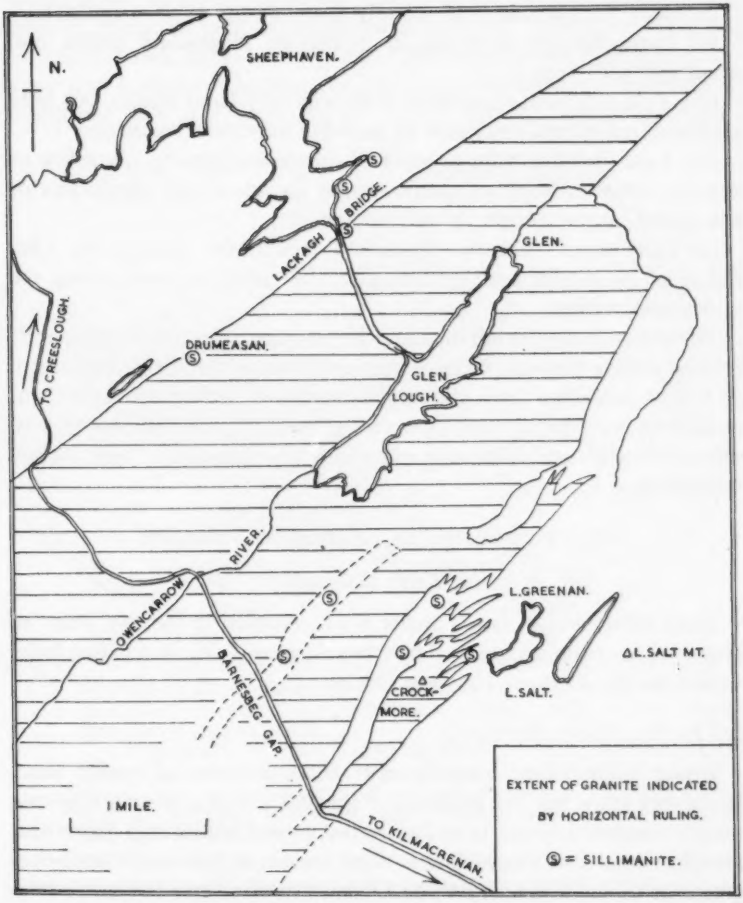
Introduction

As a result of metamorphism by the north-eastern termination of the main Donegal Granite, sillimanite appears in pelitic and semipelitic bands in Dalradian rocks near the village of Glen in northern Donegal. Here the outcrop of the granite has tapered down to a width of about two miles, so that the rocks from both contacts can be examined within a quite small area (see Text-fig. 1). Some of the sillimanite occurrences included in the scope of this paper were noted by McCall (1954, pp. 170–1) when studying the geology of the Creeslough district, which lies immediately to the north-west of the area under consideration.

The local high-grade metamorphism is a contact effect produced by the granite, to which it is spatially related; but plastic shearing has played quite as important a part as the thermal factor, and hence the Glen metasediments show the main features of regionally metamorphosed schists. Textures are schistose and all the index minerals of the higher Barrovian zones have been formed. High-grade index minerals occur only in two narrow belts of country rock bordering the contacts, and in two slices of country rock lying inside the south-eastern granite boundary; even within these limits their distribution is both sparse and sporadic. The schists are generally fine grained and sillimanite is megascopically visible only in extremely rare cases. This mineral is interesting, however, by reason of its varied textural and time relationships with the other minerals of the schists.

Text-fig. 1 demonstrates the very patchy distribution of sillimanite and the close proximity to granite of all the rocks in which it has been found. Along the north-west margin sillimanite is developed in a pelitic band to the north of Lackagh Bridge, and in coarse pelitic gneisses immediately in contact with the granite at Lackagh Bridge and farther south-west at Drumeasan. On the opposite side of the granite it occurs in two slices of metasediment running transverse to Barnesbeg

Gap in a north-east to south-west direction. One of these separates the outcrop of the granite sheet of Crockmore from the main granite area. Hand-specimens of the pelitic schists in this sedimentary slice show small flakes of muscovite, up to about an eighth of an inch in diameter,



TEXT-FIG. 1.—Outline geological map of the Glen district, Co. Donegal, showing the areas of Dalradian rocks (blank) and granite (lined), and the distribution of known sillimanite localities.

which are often set oblique to the schistosity. These become much more numerous towards the contacts of the Crockmore granite sheet and its offshoots. The field-relationship shows that the muscovite flakes result from contact action by the muscovite- and garnet-bearing granite of the Crockmore sheet, which is very slightly younger than the main granite mass.

Genetic Varieties of Sillimanite

In the sillimanite-bearing rocks taken as a whole, five types of sillimanite can be distinguished differing from each other in their mode, or time, of origin. These are :

(i) Early fibrolite derived directly from biotite ;

(ii) Early fibrolite enclosed in quartz or plagioclase grains and occurring with type i ;

(iii) Fine sillimanite prisms derived originally from biotite, but later recrystallized during enclosure by growing muscovite metacrysts ;

(iv) Late fibrolite, a by-product of muscovite growth, occurring as needles either in pools of quartz within the muscovite metacrysts or else based on the margins of the metacrysts ;

(v) Late, more coarsely crystalline, sillimanite intergrown with flakes of muscovite and occurring in the pelitic gneisses along the north-west contact.

The above five types fall into two time groups : first, early sillimanite formed during the main stage of high-grade metamorphism (types i and ii) ; and secondly, later sillimanite produced during the growth of muscovite metacrysts, and comprising both regenerated early sillimanite (type iii) and sillimanite of entirely new generation (type iv, and perhaps type v).

Description of the Sillimanite Varieties

(1) *Early Sillimanite*

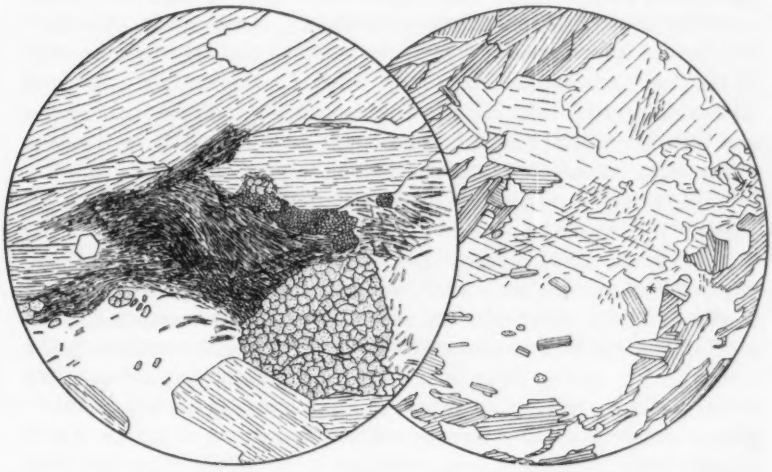
Early sillimanite of type i and ii is encountered in the two strips of country rock running across Barnesbeg Gap, and also in a pelitic band a little to the north of Lackagh Bridge.

(*a*) *In schists from Barnesbeg*

Under high-power magnification thin sections of rocks from Barnesbeg show varying amounts of acicular fibrolite set in a schistose matrix composed of quartz and a russet-coloured biotite with prominent pleochroic haloes. Plagioclase is often present as fine equidimensional grains or as small metacrysts, and muscovite in coarse flakes is sometimes an additional constituent.

Numerous sheaves and elongated lenticular mats composed of extremely fine pale-brown needles of fibrolite are developed along biotite-rich folia, parts of which they pseudomorph. In the less altered parts of these folia dense tufts of fibrolite fringe the biotite flakes or grow out from their terminations. The proximal ends of the needles are often embedded in the biotite, and so close is the association of the two minerals that it is frequently difficult to distinguish the exact course of the boundaries between them (see Text-fig. 2A).

As the ratio of fibrolite to biotite increases, so the colour of the latter becomes progressively paler. Even where biotite has apparently disappeared, tiny patches with residual pleochroism frequently persist in the interior of the dense fibrolite aggregates which have taken its place. Minute grains of magnetite which usually crowd the fibrolite areas are probably by-products of the replacement process. Whereas sillimanite has formed from, or taken the place of, biotite, it is not strictly correct to say that it replaces biotite, since many of the early formed needles project out from the biotite flakes into the surrounding groundmass.



A. B.

TEXT-FIG. 2.—*A*. Sillimanite Schist, Barnesbeg. × 65. A felted sheaf of fibrolite needles replacing biotite impinges on an area of much coarser sillimanite of granular appearance. Biotite with closely spaced cleavages ; quartz blank.

B. Muscovite-bearing Schist, Barnesbeg. × 10. A trail of short needles of recrystallized sillimanite passes through two late muscovite flakes in a direction parallel to the schistosity. Biotite, closely spaced cleavages ; muscovite, widely spaced cleavages ; plagioclase and quartz, blank.

In general the sillimanite forms extremely slender crystals, but occasionally a cluster of fibrolite needles has in its centre a number of coarser crystals which appear like anhedral grains, some of which nevertheless show a roughly diamond-shaped cross-section with a single well-marked cleavage parallel to one diagonal. Though in most cases the fibrolite needles tend to lie sub-parallel to the biotite foliation which they replace, yet in other cases they are aggregated together to form knots or plumose groupings, whilst in a small slab of sediment caught up in the Crockmore granite sheet they are arranged so as to form a typical hornfelsic texture.

In some of the thin sections described above fibrolite also occurs in quartz and plagioclase grains of the groundmass as groups of rather short needles with random orientation (type ii). Although these needles have not been formed directly from biotite there is no reason to believe they are non-contemporaneous with the fibrolite already described.

(*b*) *In schists from the north-west contact*

Comparable rocks from the north-west side of the granite resemble the Barnesbeg rocks in texture and in mineral relationships, but show a much more extensive replacement of biotite. Hence where sillimanite has formed it is present in appreciable quantity and the biotite-content of the pelite has been substantially reduced. Prismatic crystals and coarse skeletal growths of sillimanite accompany the acicular variety, and the new mineral occurs in quartz-rich folia as well as in the folia which were previously biotite-rich.

A high degree of preferred orientation of the sillimanite crystals appears to be present and to extend even to extremely fine needles enclosed in grains of quartz. Control by the pre-existing rock-fabric might suffice to explain the orientation of those crystals which replace biotite, but the corresponding arrangement of the quartz-encased needles must result from the operation of shearing movements in the country rock during sillimanite formation. Metamorphism at this contact of the granite was rather more intense and accompanied by greater deformation of the recrystallizing rocks than at the south-east contact, these differences being reflected in the nature of the resulting rocks.

2. *Late Sillimanite*

The term " late sillimanite " is used in this paper to include all sillimanite whose final crystallization was contemporaneous with the growth of muscovite metacrysts characterizing the later stages of metamorphism. It is considered that formation of the new mica was induced by the passage of late-stage potash-rich fluids from the granite into the country rocks, and in the Barnesbeg area the close spatial relation between the distribution of muscovite metacrysts and the contacts of the Crockmore granite sheet is evident. Since the growth of muscovite in these schists has left a great part of the earlier textures undestroyed, their description will precede that of the completely reconstituted pelitic gneisses of the north-west contact.

(*a*) *In schists from Barnesbeg*

The muscovite metacrysts of the Barnesbeg schists sometimes lie along the schistosity of the enclosing rock, but more often they are arranged in an entirely random fashion indicating crystallization under

static conditions. Many of the thin-sections show early biotite-derived fibrolite persisting only half a centimetre or so from the new metacrysts. It is evident that reconstitution was extremely restricted in spatial range, and in fact embraced little more than the micro-environment immediately surrounding each new muscovite flake. The great bulk of the rock material in the muscovite-bearing schists from this locality underwent little modification in this later stage of metamorphism.

The new mica appears to have formed largely at the expense of biotite, though textural evidence of such a replacement is lacking. Neighbouring biotite flakes sometimes appear to have recrystallized, perhaps with the destruction of early-formed sillimanite, since this is always absent from such flakes. Biotite also occurs as thin laminae intergrown with the muscovite or as small included flakes with varying orientations and oblique to the cleavage of their muscovite host.

Other common inclusions in the muscovite flakes are irregular pools of quartz, needles of sillimanite, minute granules of epidote, and stumpy prisms of apatite. The quartz of each pool normally consists of a single grain considerably larger than the average quartz grain of the groundmass. Presumably the pools represent recrystallized groundmass quartz which was present in excess of the amount required for muscovite formation. The apatite prisms are perhaps a legacy from former biotite.

Sillimanite enclosed in muscovite (type iii) occurs only in small quantities and tends to form trails of stumpy needles oriented roughly parallel to the schistosity of the rock, though generally oblique to the cleavage of their host (see Text.-fig. 2B). Such trails are interpreted as relics of sheaves of early fibrolite. A consideration of the small amount of sillimanite persisting in this fashion shows that much early sillimanite must have been utilized in building the plates of white mica. The modified habit of the surviving crystals indicates that they underwent recrystallization at the same time.

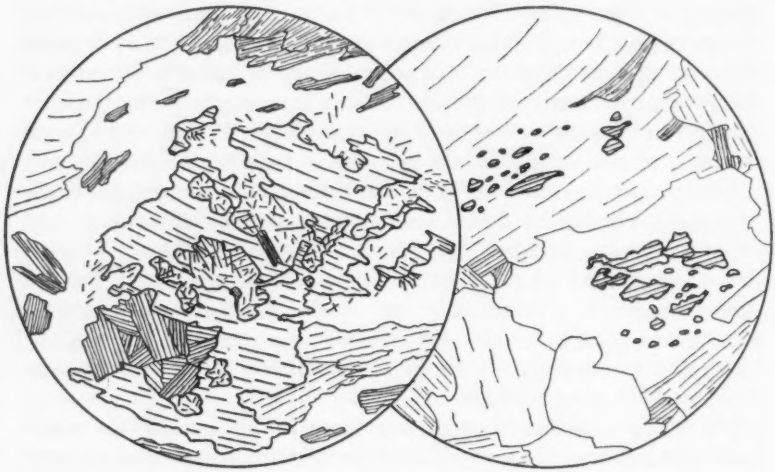
Numerous tiny sillimanite needles (type iv) crowd the pools of enclosed quartz and also quartz grains which are external, but adjacent to the new muscovite (see Text-fig. 3A). Sometimes these needles are based on the margins of the muscovite flakes, though more often they are completely isolated in the enclosing quartz. While it is possible that these particular needles may have been built of the material of former early sillimanite, it is clear that they are entirely new crystals genetically connected with the growth of the muscovite.

In addition thin sections of the same rocks sometimes show fine sillimanite needles similar in their mode of occurrence to those described above as type ii. Whether these are relict, or recrystallized, needles of early sillimanite of type ii, or whether they are of later initiation, cannot

be decided. Sometimes swarms of fine wisps of muscovite can be observed, which appear to have pseudomorphed needles of sillimanite. Such wisps are found in quartz and also in the large muscovite flakes. In the latter case replacement clearly took place subsequent to the growth of the enclosing mica.

(*b*) *In pelitic gneisses from the north-west contact*

The last setting in which sillimanite is found is provided by the pelitic gneisses which occur, as small isolated patches, in immediate



TEXT-FIG. 3.—*A*. Muscovite-bearing Schist, Barnesbeg. × 10. Needles of late sillimanite crowd quartz pools and embayments in a ragged metacryst of muscovite (outline and cleavage emphasized). Biotite, closely spaced cleavages; muscovite, widely spaced cleavages; quartz, blank.

B. Pelitic Gneiss, Drumeasan. × 10. Coarse, intergrowth of muscovite and sillimanite shown by crystallographic parallelism between sillimanite areas enclosed in two adjacent muscovite flakes. Biotite, closely spaced cleavages; muscovite, widely spaced cleavages or blank; sillimanite, heavy margins and cleavages.

contact with the granite at a number of localities along its north-western edge. These are the most coarse grained of all the pelites in the area, and have been completely reconstituted by static recrystallization during a locally developed late stage in the metamorphism. Some addition of material from the granite probably accompanied this process. The constituent minerals are (in decreasing order of abundance) biotite, muscovite, oligoclase, and quartz. In thin-section sillimanite appears only in very small amounts, if at all.

When present sillimanite nearly always has muscovite as its host. Biotite is perfectly fresh, and shows no signs of the replacement by

fibrolite so prominent in the sillimanite-schist from this contact (see Section 1*b*). Sometimes sillimanite occurs in muscovite as swarms of tiny parallel prisms, but the direction of elongation of the swarms is inconstant and without parallelism with the schistosity. This might, indeed, be expected in a rock which has been so thoroughly recrystallized.

Occurrence of sillimanite as quite coarse crystals which are intergrown with the muscovite in an irregular fashion is restricted to these gneisses. In this case the sillimanite appears in thin-section as groups of small diamond-shaped, or irregular, areas enclosed either in a single flake, or in a number of adjoining flakes of the mica (see Text-fig. 3B). Often a number of adjacent areas are optically continuous, and their perfectly fresh condition forbids their interpretation as relics of earlier complete crystals now partly replaced by the muscovite. All evidence of the mineralogy and detailed textures acquired by these rocks in the earlier main stage of the general metamorphism has been obliterated in the recrystallization which gave them their present gneissic form. Hence only the final relationship between sillimanite and muscovite, which points to simultaneous crystallization, or recrystallization, of the two minerals, can be observed.

Discussion

An examination of the literature dealing with sillimanite in metamorphic rocks shows that a close association between sillimanite and both biotite and muscovite has repeatedly been recognized. Hall (1909), Hall and Molengraaf (1925), Tattam (1929), Turner (1933), G. J. Williams (1934), L. L. Smith (1945), and J. Watson (1949) have all described fibrolite arising from boitite in a fashion similar to that exhibited in the Glen rocks. The frequency with which this genetic association recurs in a diversity of environments (in contact-metamorphosed pelites, in schlieren in gneiss, at junctions between veins and country rock, and in a migmatitic terrain) makes it likely that this is the most common paragenesis of the sillimanite of metamorphic rocks.

We have seen that many of the Glen contact rocks have the characters of regional schists. The concept of progressive metamorphism, formulated by Harker (1932, p. 208) implies that high grade regionally metamorphosed rocks have undergone " a succession of changes which follow in response to a continued rise of temperature ". When writing of the pelitic rocks of the sillimanite zone, the same author states (pp. 227–8) " in the highest grade of all cyanite gives place to sillimanite . . . It is to be remarked, however, that the sillimanite is derived in part from another source ". This putative source of sillimanite is found in the dissociation of the micas. Muscovite supposedly yields

sillimanite, and it is tentatively suggested that biotite may yield garnet. It should be noted, however, that Harker recognized the formation of sillimanite from biotite during contact metamorphism (loc. cit. supra, p. 58).

The sole record known to the present writer of the actual formation of sillimanite from kyanite is Weiss' claim (1949) to have observed pseudomorphs of sillimanite after kyanite in the Wissahickon Schists, a claim which Wyckoff (1952) was unable to confirm. Although close association of sillimanite and muscovite is common, it seems that the exact significance of their relationship is rarely obvious. A number of writers, amongst them Tattam (1929), Williams (1934), Billings (1938), Sanders (1954), and Iyengar, Pitcher and Read (1954) have described replacement of sillimanite by muscovite ; but few have suggested the existence of the converse relationship.

In many slides of the Barnesbeg schists, fibrolitization of biotite is only incipient, and it is evident that if kyanite had ever been present numerous relics of it would still survive. When fibrolite began to form these rocks were simply biotite-quartz schists. Likewise kyanite-bearing schists, which are also present along the north-west side of the granite, show no petrographic evidence of kyanite giving rise to sillimanite.

The chemical changes involved in the fibrolitization of the biotite-quartz schists consisted of dissociation of the mica with the formation of sillimanite and the expulsion of K, Mg, Fe, F, and (OH), but the iron sometimes remained as grains of magnetite. A certain amount of diffusion of the elements of which sillimanite is composed is shown by the growth of fibrolite needles outwards from biotite into the surrounding matrix, and also by the presence of the same mineral in quartz and plagioclase grains not directly associated with biotite (type ii). In extreme cases where most of the biotite has been converted to sillimanite, as in the schist near Lackagh, selective subtraction of material was doubtlessly facilitated by the contemporaneous deformation of the rock due to forcible intrusion of the granite.

Growth of the late muscovite metacrysts marks a local introduction from the granite of K, F, and (OH) which partly annulled the slightly earlier loss of certain elements and led to a complex redistribution of the material of the schists affected. Muscovite grew at the expense of biotite, of fibrolite, and probably of quartz ; some early fibrolite survived and new sillimanite was formed ; and in addition there was limited regeneration of biotite. These readjustments took place before temperatures had declined appreciably from their maximum. After a short interval replacement of sillimanite needles by muscovite pseudomorphs followed at some points in the rocks. The sillimanite-muscovite relations, when considered as a whole, seem to indicate

that both minerals were stable under the prevailing physical conditions, and that persistence, or complete obliteration, of sillimanite depended on extremely local variations in the chemical composition of the schists (i.e. on variations in the amounts of K and (OH) present).

Contrary to what might be expected, the pelitic gneisses immediately in contact with the north-west side of the granite do not show an intensification of the subtraction process which has affected the nearby sillimanite-schists ; and the extremely small sillimanite content of the pelitic gneisses seems to indicate an appreciable introduction of K from the granite. Petrographic evidence gives no clue as to whether the gneisses were first converted into sillimanite-rich rocks which only acquired their present mineral composition at a late stage, or whether excess of K inhibited sillimanite formation during the whole course of their metamorphism.

Study of the Glen contact-pelites shows that the period of high-grade metamorphism did not consist of a single stage. Two principal stages can be distinguished, the first predominantly thermoplastic in nature, the second characterized by varying degrees of recrystallization resulting from late-stage introduction of material from the granite. Variations in the amount, character, and textural relationships of sillimanite reflect the complex redistributions of the rock material which accompanied both stages of the high-grade metamorphism.

Acknowledgments

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GEOLOGY DEPARTMENT,
IMPERIAL COLLEGE,
LONDON, S.W.7.

The Kirkmabreck Granodiorite, near Creetown, South Galloway

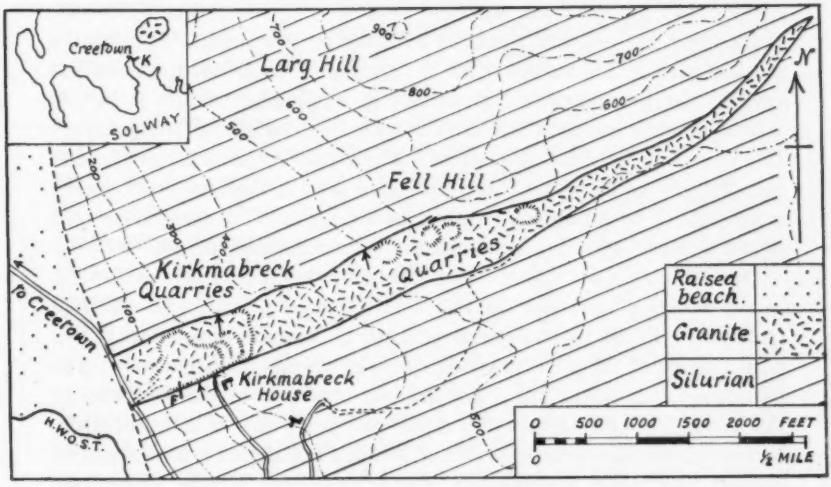
By F. G. H. Blyth

(PLATE XVI)

Abstract

The structure and flow-layering of a granodiorite dyke, as seen in extensive quarry exposures, are described and related to quarrying directions and the regional tectonic pattern.

About one and a half miles south of Creetown a dyke-like mass of grey granodiorite is intrusive into the folded Lower Silurian sediments of the district, and has produced in them a small aureole of contact metamorphism. The outcrop of the granodiorite has a maximum width of some 550 feet, and extends from the Solway coast east-north-eastwards for nearly 1½ miles, parallel to the regional strike



Text-fig. 1.—Geological sketch-map of the Kirkmabreck granodiorite. Inset shows the locality (*K*) in relation to S.W. Scotland.

of the sediments, and tapering off in that direction (Text-fig. 1). Extensive exposures are available in the Kirkmabreck quarries of the Scottish Granite Company and in Messrs. Stewart & Co.'s nearby quarries on Fell Hill, where deep excavations have been made. For many years the rock has been worked for constructional stone, and it is listed by Howe (1910, p. 415) and by Watson (1911, p. 255). It was also mentioned by Teall (1899, p. 624), but he does not give the

name of the locality. The rock is, on the whole, uniform in composition, but parts of it show an orientation structure due to the parallelism of micas, together with a marked joint pattern. These structural features, of which no record is known to the writer, form the main subject of the present paper. A similar but smaller granitic body occurs at Carsluith, a mile further south, and this also has an east-north-easterly extent parallel to the regional strike. (1-inch map : Scotland, sheet 4.)

At the Kirkmabreck Quarries, which adjoin the main road from Castle Douglas to Creetown, both walls of the granitic " dyke " are seen. The quarries consist of a series of workings, the lowest being at road level (approximately 35 feet O.D.), with successively higher levels extending eastwards into the hillside. The northern contact of the granodiorite against hornfelsed sediment is just accessible at the top of the high-level quarry face due north of Kirkmabreck House, and is marked by a dip arrow on Text-fig. 1 ; the contact dips 65° N.N.W., parallel to the bedding of the hornfelses. The southern contact is exposed near floor level on the south side of the lowest quarry, about 650 feet in from the road, and dips 55 to 60° N.N.W. ; this south wall, from which granodiorite has been removed by quarrying, is in effect the contact surface and strikes N.60° E., again parallel to bedding. At this contact the granodiorite has developed a selvedge, about ½ inch in width, of finer textured and darker rock, the darkening being mainly due to an increased biotite content. The granodiorite sends small veins into the sediments here. Veins are again seen in the high-level quarry north-east of Kirkmabreck House ; at this higher level the whole width of the granitic mass can be inspected, over a distance of 480 feet.

Other exposures of the northern contact occur higher up the hillside, in the group of quarries on Fell Hill (referred to below as the Fell Quarries). In the most westerly of these the contact dips 55° N.N.W., with a strike of 68° ; the granodiorite is of normal texture nearly up to its margin, but again has a fine-grained border. At the west end of this quarry, fracture-cleavage is developed in a belt 5 to 6 inches wide, between unbroken granodiorite and bedded hornfelses ; movement of normal fault type is indicated, the granodiorite having moved up relative to the hanging wall of sediment—probably a post-emplacement movement. Standing water fills this quarry to a level of nearly 500 O.D. In the most easterly of the Fell Quarries, hornfelses are again seen high up on the northern face, but any exposures of the southern contact are obscured by tip-heaps.

The granodiorite as a whole thus has the form of a dyke, trending nearly E.N.E. and inclined at about 60° N.N.W. ; it thins out towards the E.N.E. The western termination of the dyke, and of that at Carsluith, is not exposed ; it may taper off, or it may be cut off by a north-south fault lying along the Cree estuary.

Petrography of the Granodiorite

In average mineral composition, measured by point-counter on several thin sections, the quarry-rock from Kirkmabreck is made up of plagioclase (52), quartz (31), microcline (7), and mica (mainly biotite, 10 per cent). The rock has a moderately coarse texture, the average size of plagioclases being 1 to 1½ mm., and is not porphyritic. The plagioclase (oligoclase-andesine) shows a tendency to euhedral outlines against quartz ; crystals are zoned, with cloudy centres and clear, more albitic borders. Much secondary white mica, often oriented along cleavages, is developed in the cores of the crystals, which then resemble those described from porphyrite dykes found in the neighbouring ground to the east (Blyth, 1949, p. 398). The microcline or microcline-perthite occurs in large plates nearly comparable in size to the plagioclase.

The quartz forms either large crystals which show strain polarization, or granular aggregates some 2 mm. across which result from the breakdown of large crystals under strain. The prevalence of strained quartz, seen in all the thin sections of the granodiorite which have been examined, is a distinctive feature of the rock. A small amount of quartz replaces some of the feldspar.

The mica is mainly biotite, in crystals up to about 1 mm. in length, with pleochroism from reddish-brown to straw yellow. A little muscovite is intergrown with the biotite, or occurs as sheafs of small flakes, but it amounts to only a small fraction of the total mica. Some biotite shows incipient alteration to chlorite. No hornblende has been observed. Accessories include apatite, granular sphene, zircon, and in some sections a little interstitial calcite. Magnetite is rare.

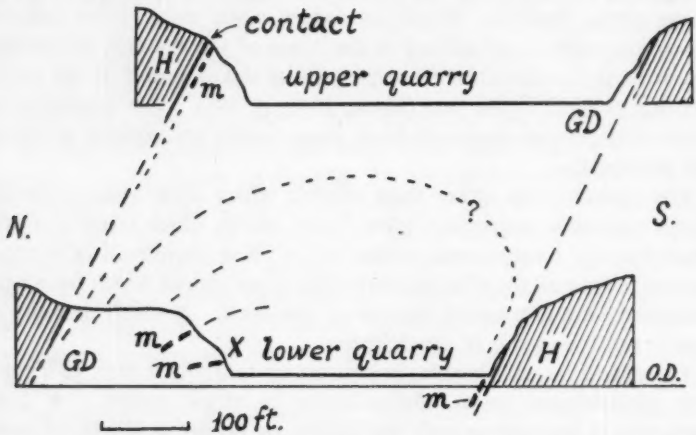
At the northern contact of the dyke the granodiorite becomes a porphyrite, with an andesitic rather than microdioritic texture. Small, beautifully zoned porphyritic plagioclases, generally less than 0·5 mm. in length, together with aggregates of minute biotite flakes, are set in a microcrystalline matrix of feldspar, quartz, and biotite. The biotite aggregates are elongated and have a marked parallel orientation, as described below ; they are probably derived from the hornfelses. This fine-grained modification of the granodiorite is not developed to the same extent at the southern margin of the mass.

Flow Layering

A definite but rather feeble lineation is visible on many joint surfaces in the granodiorite as a result of the parallel orientation of a proportion of the biotites in the rock. At the northern contact of the dyke, seen high up in the wall of the upper quarry at Kirkmabreck as already described, the parallelism of the micas is more marked : they lie in

planes which dip at 50° and follow the contact. This orientation of the biotites near the bounding wall of the dyke is attributed to magmatic flow during intrusion : the micas delineate flow-layers.

Oriented specimens were collected near the foot of the north wall of the low-level quarry, some 200 yards in from the road, and also near the south wall of the same quarry. The specimens from the north side of the quarry have biotites which lie with their flat surfaces in planes dipping north at about 15°. In one slice cut approximately parallel to



TEXT-FIG. 2.—Diagrammatic sections through the Kirkmabreck Quarries, projected along the strike of dyke, to show probable flow-layers in granodiorite (*GD*) and contacts with hornfels (*H*). *m* = observed oriented micas.

this plane, over 60 per cent of the micas occur as basal sections. The point from which the specimen was taken must be at least 150 feet (horizontally) from the wall of the dyke, assuming the dip of the latter is constant (Text-fig. 2). Forty feet or so south of this point, on a projecting corner of the unquarried granodiorite (*X* in Text-fig. 2), a smaller proportion of the biotites in the rock lie with their flat surfaces nearly horizontal. The flow-layers are evidently " turning over " from a 50° dip near the contact to a horizontal attitude some 200 feet away from it, thus forming part of an " arch of flow-layers " (Balk, 1948, p. 113). The possible shape of this arch as inferred from several observations is shown in the figure. The horizontality of the flow-layers in the central part of the granodiorite intrusion, where quarrying has been carried on to the greatest extent, probably explains the attitude of the easy parting direction in the rock, as discussed below.

Specimens from the south side of the same quarry at Kirkmabreck showed a much less marked degree of orientation of the flow elements, not more than 20 per cent of the micas having a parallel arrangement

in planes dipping north parallel to the wall of the dyke. The rock here also retains its normal texture almost up to the contact, as already described, with only a selvedge of modified granodiorite adjacent to the wall-rock.

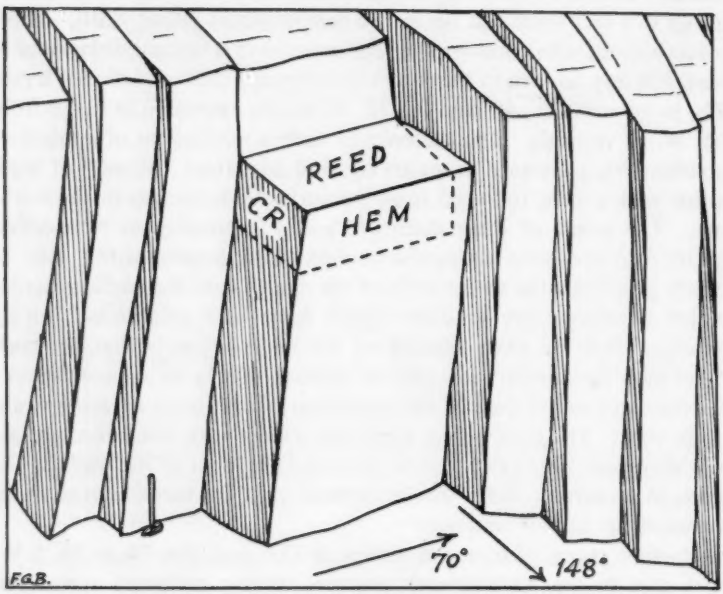
The Joint System

Dominating the structures in the rocks of both the Kirkmabreck and Fell Quarries is the pattern formed by a series of north-north-westerly joints, which break the granodiorite into nearly vertical slices about two feet thick but up to five feet in places (Plate XVI). These joints extend, with considerable regularity, over a vertical distance of at least 600 feet, as seen in the several quarry exposures at different levels. The joints strike at 148° and dip 70° W. on the average (dip varies from 65° W. to vertical). The presence of such a marked set of joints is of considerable assistance in quarrying, and particular " slices " of high-grade rock can be followed from the surface downwards in the working. The origin of these dominant joints is possibly to be ascribed partly to contraction during the cooling of the granodiorite : they lie nearly perpendicular to the walls of the dyke, i.e. to the surfaces across which a temperature gradient would have been established during cooling. But the close spacing of the joints suggests that the rock mass may have been deformed by stresses during its consolidation ; the fractures would then be the expression of maximum shearing strain in the rock. The joint system resembles a large-scale fracture-cleavage, and may have been produced by relative movement of the walls of the dyke, in a manner similar to the formation of fracture-cleavage in an incompetent bed of sediment.

Another group of fractures strikes at 133° and dips 70 to 75° S.W. It is seen in the Kirkmabreck quarries cutting obliquely across the 148° joints. These fractures carry slickensides which pitch 30° to 35° N.W. ; they are therefore small faults, with a larger horizontal than vertical component of movement, and in some instances the movement can be shown to be right-handed. One of the fractures, in the north wall of the high-level quarry, is filled with quartz and shows traces of copper mineralization (an old adit high up in the south wall of the lower quarry was used in the past to gain access to a copper lode). Small north-south faults with mainly horizontal movement are also present ; one crosses the low-level quarry about 150 yards east of the main road, and brings hornfelsed sediments against granodiorite over a short distance in the south wall of the quarry. There is thus evidence that small post-emplacement movements affected the rocks.

A somewhat less prominent set of joints traverses the granodiorite in the 70° direction (i.e. parallel to the contact) ; these joints are smooth, vertical or dipping steeply south, and are well seen in the

north-east corner of the upper quarry. One such joint-surface here, exposed for some 25 feet, has small vughs developed along irregular cracks in the granodiorite ; many micas, muscovite as well as biotite, adhere to the face, which probably represents a channel along which volatile constituents passed through the rock during consolidation. The face ends at one of the slickensided joints (133°) noted above. Other parallel surfaces striking at 70° continue in the granodiorite in this northern wall of the upper quarry, at the top of which the northerly



TEXT-FIG. 3.—Directions of easy parting (*reed* and *hem*) in relation to main joints (diagrammatic). *CR* = *cross-reed.* Bearings shown by numbered arrows.

dipping contact is exposed. The 70° joints therefore make a vertical acute angle with the contact, unless the latter changes its dip below the point at which it is exposed. Other joints observed trend 23 to 30° and are vertical.

In the Fells Quarries, all the joints described above are present, the 148° set being always the most prominent. In the highest (most easterly) quarry, several aplite veins about 6 inches wide traverse the granodiorite in a nearly east-west direction, dipping 60° north. In the most westerly quarry of the group, slickensides on a vertical face at 163° pitch 30° S., and indicate some horizontal movement on a small fault.

Quarrying Directions

In the quarrying operations, advantage is taken of the 148° and 70° joint directions in extracting the stone. The easiest parting, locally called the *reed*, lies nearly horizontally (Text-fig. 3) and, in general, is not readily apparent in the rock. As already stated, it is probably related to the flow orientation of biotites, which in the central part of the dyke lie in horizontal surfaces but are not conspicuous. The vertical joint direction at 70° constitutes another easy parting, called the *hem* ; reed and hem together with a third direction of splitting, the *cross-reed*, yield roughly rectangular blocks of stone. The cross-reed corresponds to the " hard-way " of other localities, and cuts across the flow-layers in the rock. I am indebted to Mr. J. A. MacGregor, of the Scottish Granite Company Ltd., for this information concerning the directions of working in the granodiorite, and also for permission to inspect the quarries freely.

Conclusions

Certain inferences may be drawn from this study. The directions of jointing or fracturing in the granodiorite, and the suggested mechanism for the formation of a fracture-cleavage on a large scale by relative horizontal movement of the walls of the dyke, correspond closely to similar features found in the sheared porphyrite dykes which lie to the east of the locality under discussion (Blyth, *loc. cit.*). The directions are regional directions. The porphyrite dykes were shown to be fault-intrusions, and the granodiorite may also be a fault-intrusion which, after its emplacement, partook of movements which affected the area during a late phase of the Caledonian orogeny.

The filling of the Kirkmabreck dyke probably took place in a mainly upward direction. A similar mechanism has been postulated to explain the vertical southern margin of the Cairnsmore of Fleet granite (Blyth, 1954), which has a similar strike to the Kirkmabreck dyke and follows a parallel line a few miles to the north-east. It was suggested that the marginal part of the granite may have become emplaced by the coalescence of several parallel dyke-like intrusions (*loc. cit.*, p. 246). The Kirkmabreck granodiorite possibly provides a small-scale example of this mechanism, the magma in both instances having been generated at depth and having risen along steeply dipping or vertical planes of weakness in the closely folded sediments of this part of the Southern Uplands.

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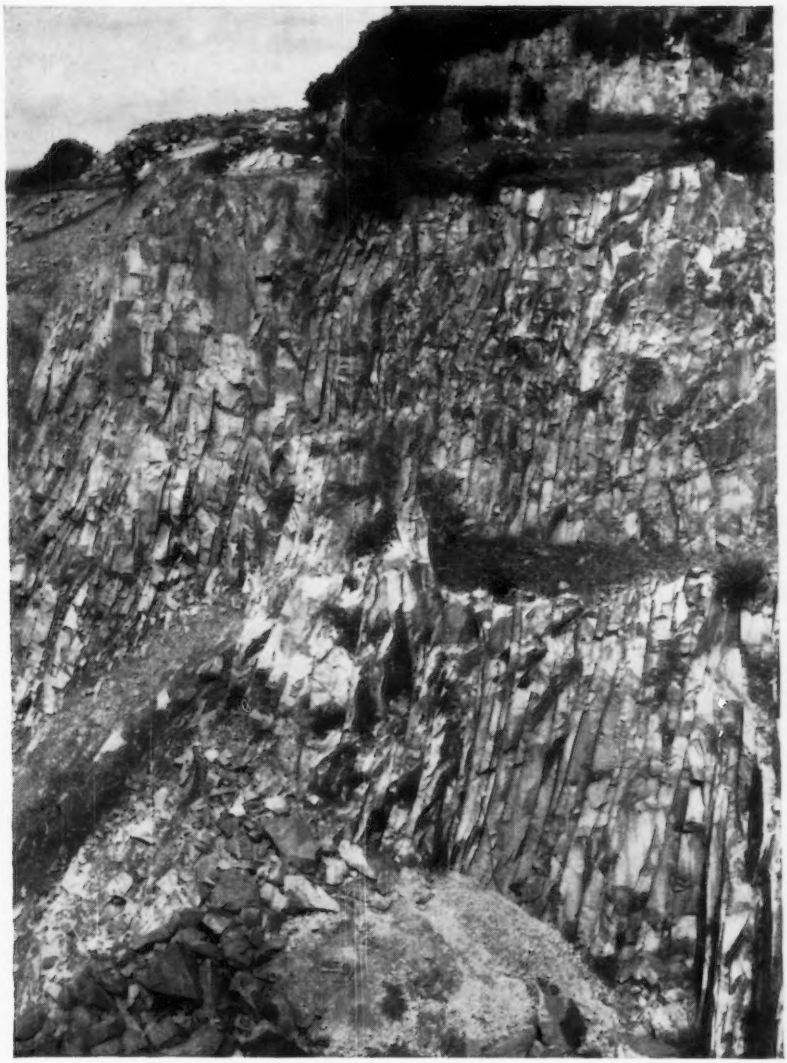
GEOLOGY DEPARTMENT
IMPERIAL COLLEGE,
LONDON, S.W.7.

EXPLANATION OF PLATE XVI

View of upper part of the Kirkmabreck Quarry, looking towards the north wall; the nearly vertical joint "slices" cross the granodiorite in a north-north-westerly direction. Upper level of working seen at top right.



JOINTS IN KIRKMAEREECK GRANODIORITE.

The Origin of the Girvan-Ballantrae Beerbachites

By T. W. Bloxam

(PLATE XVII)

Abstract

The mode of occurrence and petrography of a group of pyroxene-hornblende-granulites (beerbachites), found only in association with serpentinized peridotite, are described. The beerbachites are regarded as contact altered lavas and basic intrusions which formed the lower levels of a thick series of Ordovician volcanics. Peridotite, intruded into the base of the volcanic pile, disrupted and thermally metamosphosed the lavas and intrusions which appear as inclusions in the serpentinized peridotite. The present contacts between serpentinite and adjacent unmetamorphosed rocks are secondary in origin.

I. Introduction and Field Relationships

A VARIED assemblage of metamorphic rocks occurs in the Girvan-Ballantrae area associated with Ordovician spilitic lavas, basic and ultrabasic intrusives. Some owe their origin to the effects of shearing associated with large dislocations and are typically albite-epidote-chloride schists and albite-epidote-amphibole schists, derived from the basic lavas and intrusions. Locally these schists pass into garnet-amphibolites (Teall, 1899, pp. 456–8; Anderson, 1936, pp. 539–542). Balsillie recorded the occurrence of glaucophane-schist and an eclogitic rock from Knockormal (1937, pp. 29–32).

The origin of a series of pyroxene-hornblende-granulites (beerbachites), found only in the southern belt of serpentinite (Text-fig. 1) has been the subject of much controversy. However, no detailed account of their distribution and mineralogy has been published.

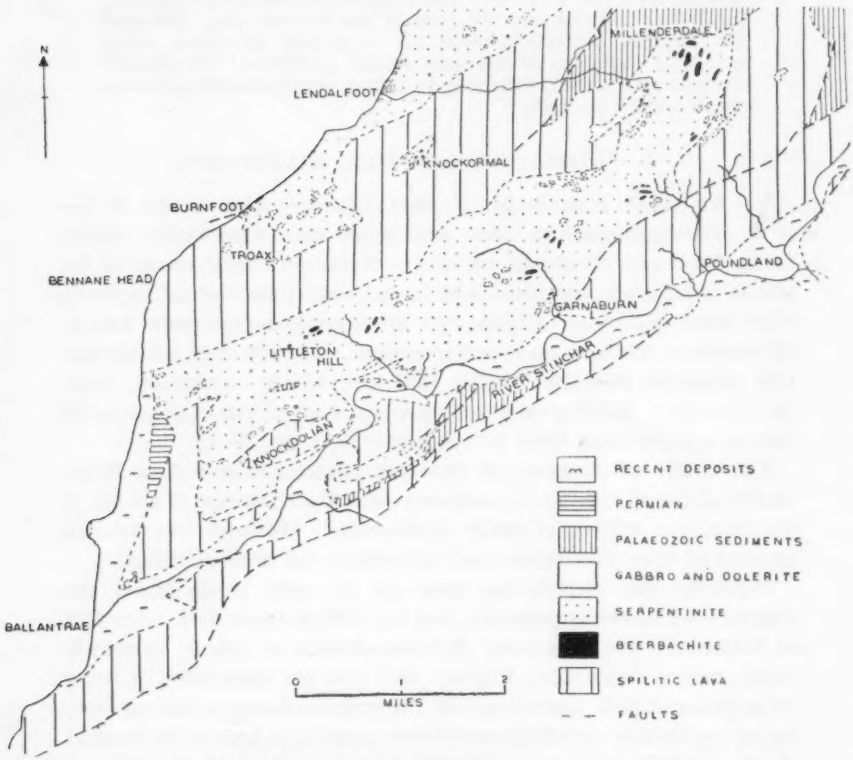
Typically, the beerbachites crop out as small knolls above the general level of the serpentinite, and are seldom more than a few feet in diameter. They are deep chocolate-brown in colour, extremely tough, with a sharp flinty fracture, and give the impression of being strongly hornfelsed. Junctions with the serpentinite are rarely exposed, but at one locality on Littleton Hill the contact is a plane of movement along which the marginal serpentinite is sheared, passing within an inch or two into serpentinized harzburgite identical to that occurring throughout the area.

Near the head of Poundland Burn an outcrop of beerbachite has a dyke-like form relative to the surrounding serpentinite. The exposure is 20 feet wide and forms a marked ridge running approximately E–W for a distance of some 150 yards. In this same locality many isolated knolls of beerbachite are roughly aligned, suggesting that they are parts of a once continuous body.

Much larger masses of flasered gabbro passing into beerbachite occur in the region of Millenderdale (Text-fig. 1).

II. Previous Work

Teall was the first to describe the " granulitic rocks " at Littleton Hill, regarding them as metamorphosed " diabase-porphyrite lavas ", possibly contact altered by the peridotite (*op. cit.*, pp. 479–480). On this basis, all other occurrences of beerbachite in the area are indicated as " contact altered spilitic lavas " on the Geological Survey, Sheet 7.



Text-fig. 1.—Distribution of beerbachite in the southern part of the Girvan-Ballantrae complex.

The same locality was examined by Tyrrell (1909) who came to the conclusion that the beerbachites were originally intrusive basic rocks (gabbros and dolerites), overwhelmed and thermally metamorphosed by the peridotite, now serpentinized.

Opposed to the " xenolithic hypothesis " of Teall and Tyrrell is the " intrusive hypothesis " of Balsillie (1932, pp. 122–3). He considered

that the beerbachites were derived from basic rocks intruded into " . . . rifts in the peridotite, which tending to close up on the still hot rock bodies, resulted in their reconstitution." An intrusive origin is also favoured by Beveridge (1950, pp. 61–3) and Bailey and McCallien (1952, pp. 30–1).

The adherents to an intrusive origin for the beerbachites admit that: (*a*) there is no satisfactory evidence of intrusion into the ultrabasics; (*b*) thermal effects from some unknown source are the dominant factor in their metamorphism, and (*c*) a comparable degree of metamorphism does occur in the adjacent serpentinite. The latter point Balsillie regarded as indicative of intrusion and metamorphism of the beerbachites prior to serpentinization; thermal metamorphism having had no effect on the original peridotite. However in 1937 (pp. 23–4) Balsillie, influenced by the association of glaucophane-schist with the serpentinite, speaks of " contact thermal or regional metamorphism " of the whole southern belt of ultrabasics and associated intrusives.

Bailey and McCallien attempt to solve the difficulty by postulating re-heating and metamorphism of basic intrusions by later invasions of gabbro and dolerite which " . . . sometimes re-used channels partially blocked by products of its own earlier crystalization." In some localities unmetamorphosed basic rocks do crop out near the beerbachites, but there are places (e.g. Littleton Hill) where no later basic intrusions can be seen anywhere near. Point (*c*) also remains unexplained.

III. Petrology of the Beerbachites

The rocks are typified, both in their mode of occurrence and mineralogy, by exposures at Poundland Burn, Littleton Hill, Knockormal and Millenderdale.

(i) *Poundland Burn.*—The central portion of the dyke-like body (p. 329) is an ophitic gabbro, but the original intergranular boundaries appear blurred in hand specimens.

In thin sections the original diallagic pyroxene is in the process of replacement by granular crystals of a clear diopsidic variety (Plate XVII, 1). This is exhibited along the cleavages and margins of the diallage which shows no signs of mechanical deformation. The diallage becomes strongly schillered with rod-like ore resulting from the exsolution and oxidation of iron and possibly titania. The optical properties of the two pyroxenes are given in Table 1.

Table 1.

	Primary (diallagic) pyroxene	*Secondary pyroxene*
Ny . . .	1·711	1·691
Nz-Nx. . .	0·025	0·028
Z : c . . .	46°	41°–42·5°
2Vz . . .	59°	53°

Amphibolization of both the original and secondary pyroxene is general. The amphibole is a strongly pleochroic, reddish-brown variety and appears as anhedral patches and streaks along the margins and cleavages of the primary pyroxene. Where it replaces the secondary pyroxene it is more euhedral and may build large crystals, often associated with irregular grains of ore. The amphibole also develops from dispersed chloritic material of doubtful origin. An analysis and optical data for the amphibole, which is ubiquitous to the beerbachites throughout the area, is given in Table 2. It corresponds closely to kaersutite.

TABLE 2.

	1		2			3		4
SiO_2	40·26	Si	5·978		8·000	39·20		41·46
TiO_2	3·38	Al	2·219	2·022		4·88		5·70
				0·269				
Al_2O_3	13·09					12·25		14·24
Fe_2O_3	0·49	Ti	0·376		4·965	2·97		3·32
FeO	11·77	Fe^{III}	0·053			10·49		5·70
MnO	0·21	Fe^{II}	1·462			0·17		0·08
MgO	12·56	Mn	0·026			11·57		13·68
CaO	12·48	Mg	2·779			12·59		11·62
Na_2O	2·18					2·68		2·29
K_2O	0·79	Ca	1·986		2·758	0·88		1·72
H_2O^+	2·34	Na	0·624			1·17		0·12
H_2O^-	0·16	K	0·148			0·02		—
P_2O_5	0·42	H	2·318			—		—
F	—					nil		0·42
	100·13					99·87		100·35
							Less 0 for F	0·18
								100·17
S.G.	3·20					3·23		3·215

X	Pale yellow/colourless	Pale yellow/brown	Light yellow
Y	Brown	Medium brown	Red-brown
Z	Deep reddish-brown	Dark brown	Dark red-brown
Nx	—	1·681	1·670
Ny	1·668	—	1·692
Nz	—	1·700	1·701
Nz-Nx	0·032	0·019	0·031
Z : c	16°	13°–15°	8°–10°
2Vx	74°	82°	81°

1. Brown Hornblende, Poundland Burn (Anal, W. H. Herdsman).
2. „ „ „ „ „ Metal atoms to 24 (O, OH).
3. Kaersutite, Skaergaard dyke. Vincent, 1953, p. 34.
4. „ Boulder Dam. Campbell and Schenk, 1950, p. 684.

Large olivine crystals with a well-developed prismatic habit and $2Vx=80°$ are sporadically developed. These crystals have also undergone marginal recrystallization into small granulitic olivine.

The original plagioclase has entirely recrystallized to a granular plagioclase, Ab_{48} An_{52} in composition. Frequently granular aggregates preserve the form of original plagioclase, exhibiting ophitic relationships to the primary pyroxene.

Towards the margins with the serpentinite the rock becomes finer in grain and porphyritic in texture. The groundmass is an even-textured aggregate of granular secondary pyroxene, recrystallized plagioclase, brown hornblende and ore. The outlines of large turbid primary plagioclase phenocrysts remain, although they are nearly all recrystallized to aggregates of secondary calcic plagioclase (Plate XVII, 2).

Veins of prehnite, carbonate, and talc occur in this, and most of the other beerbachites examined. These minerals are related to later hydrothermal activity, including serpentinization of the peridotite (Bloxam, 1954, pp. 525–528).

(ii) *Littleton Hill.*—Several textural varieties of beerbachite occur in this locality and may be divided into non-porphyritic and porphyritic (c.f. MacGregor, 1931, plate xxii). Occasionally rather schistose types are developed, but they are quite subordinate in amount. There is no ordered arrangement of the various types and none of them show signs of original chilled margins towards the adjacent serpentinite.

Mineralogically the rocks are identical with those from Poundland Burn. Recrystallization is almost complete however, the texture being saccharoidal and typically beerbachitic (Plate XVII, 3). In addition to granular pyroxene, brown hornblende and plagioclase (Ab_{46} An_{54}), occasional crystals of hypersthene may be present, together with a foxy-red biotite, typical of hornfels. When relict textures remain they are doleritic.

The porphyritic types are the most interesting although they are confined to one small outcrop. Large euhedral crystals of plagioclase, often flow-aligned, are set in a matrix of brown hornblende and granular plagioclase (Plate XVII, 4 and 5). The plagioclase phenocrysts are primary (Ab_{50} An_{50}) and show a strong thermal clouding. Their margins are often "scalloped" (Schwartz, 1924, p. 116), and the crystals may be invaded and sieved by constituents of the groundmass. Clear plagioclase in the groundmass has recrystallized (Ab_{48} An_{52}) yet retains the outlines of original small primary plagioclase laths exhibiting marked flow-banding. Pyroxene of secondary formation rarely occurs in the rock, but dusty remnants of primary pyroxene in glomeroporphyritic aggregates are common. The secondary pyroxene has probably been entirely replaced by the granular brown hornblende (MacGregor, *op. cit.*, p. 512).

This is probably the same outcrop from which Teall described the metamorphosed "diabase-porphyrite lava" (*op. cit.*, plate xx, 1). Later investigators apparently abandoned a possible extrusive origin for some of the beerbachites, attention being drawn more readily to the abundant outcrops of beerbachite clearly derived from intrusive gabbros and dolerites.

It is concluded that the porphyritic beerbachite may well be an altered porphyritic lava, strongly resembling unaltered porphyritic lavas exposed at Knockdolian Hill, Stockenray Bay and elsewhere in the area (Balsillie, 1932, p. 110). Photomicrographs of unmetamorphosed lavas in the district, all very similar in texture to the porphyritic beerbacjite, are given by Teall (*op. cit.*, plate xix, 2) and Balsillie (1932, plate vi, 4).

Occasionally schistose beerbachites are developed at Littleton Hill. The schistosity shows no uniform trend from one outcrop to the next, and in many cases is only obvious under the microscope. The dynamic effects are not connected with deformation along the present margins of the serpentinite since the latter never exhibits a comparable degree of shearing.

(iii) *Knockormal.*—Small outcrops of beerbachite, glaucophane-schist and an eclogitic rock are associated with a lenticular mass of serpentinite near the farm at Knockormal. The beerbachites here differ little from the non-porphyritic varieties at Littleton Hill (Plate XVII, 3), and are clearly contact altered dolerites.

(iv) *Millenderdale.*—Several large masses of granulitized gabbro at this locality were briefly described by Teall (*op. cit.*, pp. 477–8). The central portion of one prominent knoll is an exceedingly coarse ophitic gabbro. Thin sections show that, like the rock at Poundland Burn, large plates of strongly schillered diallage are partly recrystallized to granular diopsidic pyroxene. Towards the margins with the serpentinite the rock becomes increasingly flasered. Deep reddish-brown hornblende, similar to that already described, replaces both kinds of pyroxene. Ophitic plagioclase is recrystallized to a granular calcic variety which is frequently saussuritized. Eventually a saccharoidal beerbachite is developed which, though clearly derived from the flasered gabbro, is not itself schistose (Plate XVII, 6). Sometimes these beerbachitic bands cut across the flaser planes in the gabbro. Teall speaks of these bands as "... irregular veins of dykes of fine grained igneous rock, which, though they cut the foliated masses, are not themselves schistose" (*op. cit.*, p. 477).

It is concluded that, since the beerbachite is derived from the flasered gabbro, the intrusive relationships of the beerbachite are the result of partial mobilization of the gabbro in the formation of the beerbachitic material.

IV. Analyses

In addition to an analysis published by Balsillie (1932, p. 122), further analyses of the Girvan-Ballantrae beerbachites were made by Beveridge (*op. cit.*). These analyses are given in Table 3. For purposes of comparison, analyses of an albite-diabase and spilite, both from the same area, are included in Table 3.

There is a close similarity between all the analyses, in particular the rather high TiO_2 and Na_2O; a feature characteristic of basic rocks in spilite-serpentinite associations.

Table 3.

	1	2	3	4	5
SiO_2	45·70	43·84	51·03	47·50	47·01
TiO_2	2·60	2·32	3·04	4·49	2·94
Cr_2O_3	0·14	—	—	—	—
Al_2O_3	13·60	15·35	13·69	12·39	18·07
Fe_2O_3	2·32	2·09	2·06	4·17	5·15
FeO	9·47	10·66	12·02	10·60	5·20
MnO	—	0·21	0·22	—	0·10
MgO	10·53	8·26	4·55	4·27	3·17
CaO	10·08	8·02	7·46	7·37	10·00
Na_2O	3·74	4·09	4·23	4·88	3·46
K_2O	0·56	1·42	0·50	0·89	0·63
H_2O^+	1·19	2·63	0·82	2·62	1·11
H_2O^-	0·01	0·37	0·28	0·20	2·38
P_2O_5	0·26	0·58	0·31	0·78	0·45
FeS_2	0·08	—	—	0·02	0·14
CO_2	—	—	—	—	0·16
	100·28	99·84	100·21	100·18	99·97

1. Beerbachite, Littleton Hill. Balsillie, 1932, p. 122.
2. „ „ „ Beveridge, 1950, p. 92.
3. „ Garnaburn. „ „ p. 103.
4. Albite-diabase, Lendalfoot. Balsillie, *op.cit.*, p. 126.
5. Spilite, Stockenray Bay. „ „ p. 111.

V. Origin of the Beerbachites

It is concluded that the beerbachites were derived from a series of basic lavas and intrusions, similar chemically and texturally to unaltered lavas and intrusions at present exposed in the area. The peridotite is regarded as originally intrusive into the base of a thick series of lavas and associated intrusives which were disrupted and suffered high temperature contact metamorphism. Bodies of beerbachite with a dyke-like form and exhibiting chilled margins towards the serpentinite antedate the intrusion of the peridotite which probably ascended along channels occupied by the earlier basic dykes (Ransome, 1894, pp. 230–1).

The association of beerbachite with glaucophane-schist at Knockormal suggests that the latter may also be inclusions in the serpentinite.

The present junctions between serpentinite containing the metamorphosed inclusions and adjacent unmetamorphosed rocks are lines of considerable movement. The secondary " cold " intrusion of the serpentinized peridotite to higher levels may also have occurred (Taliaferro, 1943, p. 154, 169).

VI. Acknowledgments

The writer expresses his thanks to Professor T. Neville George and Dr. B. C. King for their help and criticism. Thanks are also accorded to Professor A. Holmes and the University of Glasgow.

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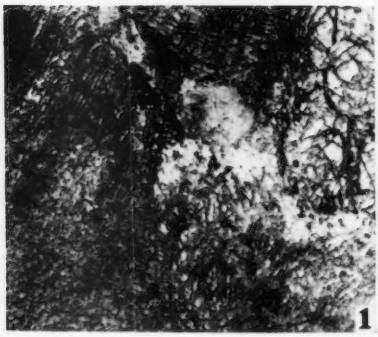
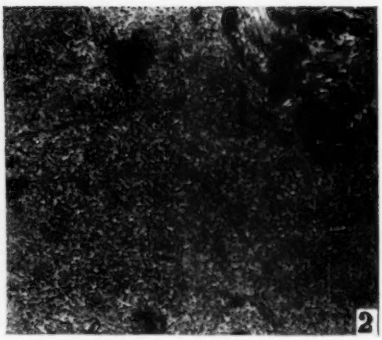
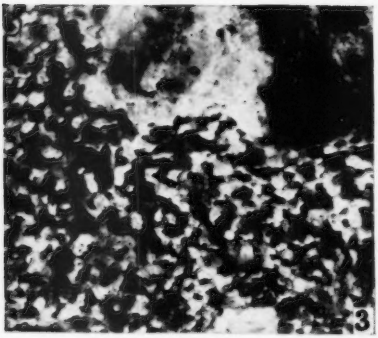
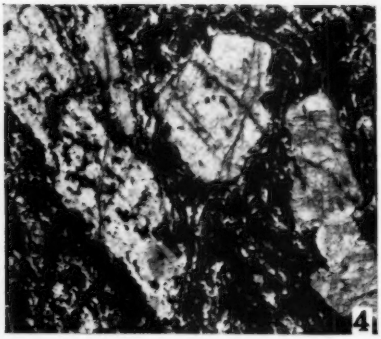
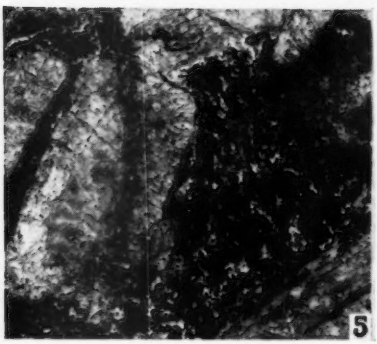
Dept. of Geological Sciences,
University of California,
Berkeley, 4,
California, U.S.A.

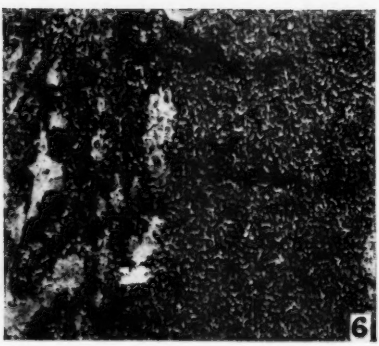












GIRVAN–BALLANTRAE BEERBACHITE AND GABBRO.

EXPLANATION OF PLATE XVII

1.—Altered gabbro, Poundland Burn. Large plates of fibrous diallage partly recrystallized to secondary diopsidic pyroxene. Deep brown hornblende replaces both pyroxenes. Large olivine crystal (top right) marginally recrystallized. Ordinary light ×20.

2.—Fine-grained marginal phase of altered gabbro. Relict plagioclase phenocrysts in a dense recrystallized matrix of secondary pyroxene, plagioclase, and brown hornblende. Ordinary light × 20.

3.—Beerbachite, Littleton Hill. Rock composed largely of granular plagioclase and brown hornblende. Partially recrystallized plagioclase phenocrysts at top right. Ordinary light × 20.

4, 5.—Porphyritic beerbachite, Littleton Hill, Flow-aligned, thermally clouded, plagioclase phenocrysts in a recrystallized matrix of brown hornblende and plagioclase; the latter exhibiting relict flow-banding. Ordinary light × 20.

6.—Flasered gabbro with deformed and partially recrystallized plagioclase augen, passing into typical granular beerbachite. Millenderdale. Ordinary light × 20.

Miocene Beds of the East African Coast [1]

By F. E. Eames and P. E. Kent

Abstract

New discoveries of marine Lower Miocene beds on the Tanganyika and Kenya coasts and on the adjacent larger islands are briefly described, and the palaeontological evidence establishing their age outlined. Previous palaeontological evidence has had to be reassessed, a considerable number of supposed Pliocene index-fossils having been found to occur in the Lower Miocene. The "*Chlamys werthi* Beds" are shown to include strata of both Lower Miocene and Pliocene ages.

Distribution

EXAMINATION of the East African coast in the course of exploration for oil by a joint BP and Shell organization has resulted in the discovery that marine Miocene beds outcrop much more extensively than has previously been supposed. J. W. Gregory (1921) had compiled records of Miocene beds in Kenya between Mombasa and Malindi; G. M. Stockley (1927) described the Miocene of Pemba Island, and other small outcrops were known on the Tanganyika coast, especially between Lindi and Mikindani. Zanzibar Island was regarded as of Pliocene and younger rocks, Mafia Island of post-Pliocene sediments only.

Recent field work, in conjunction with detailed palaeontological analysis of the marine sediments has shown that the Miocene occurs extensively on Zanzibar Island; that Mafia Island, apart from a Pleistocene fringing reef on the east coast, consists almost entirely of Miocene beneath the mantle of superficial sand; that Miocene outcrops occur on the mainland near Dar-es-Salaam and Kilwa; that Miocene beds outcrop in Lindi Bay in the south; and that Miocene faunas occur near Mombasa and Kilifi Creek in Kenya. These beds are assigned to the Lower Miocene (Aquitanian and Burdigalian stages).

Stratigraphy

(a) *Zanzibar.*—The first recognition of limestones with Miocene foraminifera was at Makunduchi in the south-east of the island, where partly haematitized limestones with an impersistent mantle of red clay appear as an inlier in the Quaternary coral terrace country; here, a

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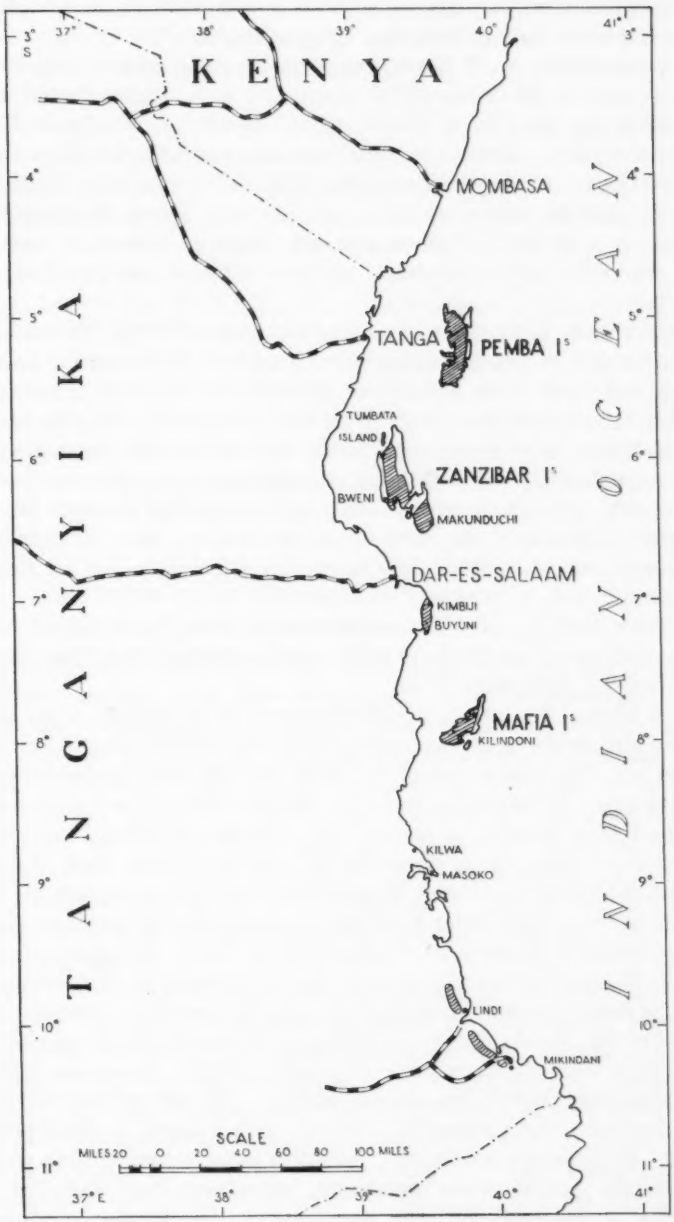

TEXT-FIG. 1.—Map of part of East Africa showing some Miocene outcrops.

Miocene species of *Operculinella* is a characteristic form, and is accompanied by *Lepidocyclina* (*Nephrolepidina*).

Subsequently, A. P. Terris found *Lepidocyclina* in a road-metal pit at Tunguu in the centre of the island, and R. E. Linton found richly fossiliferous beds in a stream section at Mtoni, 3 miles N.E. of Zanzibar City. Further work on the palaeontology has shown that a large part of the " *Chlamys werthi* Beds " of Stockley (including much material with this marker species) is of Lower Miocene date. Beds of undoubted Pliocene age are, however, known to outcrop in the north-east at Tumbatu, and near Mbweni, south of Zanzibar City.

(b) *Dar-es-Salaam district.*—Exposures in sea cliffs on the headland 20 miles S.E. of Dar-es-Salaam were found by R.B. Holroyd to include grey and green sands and silts of Mio-Pliocene date which had been tilted before deposition of the high level raised reefs. Further south near Buyuni were richly fossiliferous limestones, with pavements of *Ostrea* (*Lopha*) and beds rich in *Chlamys* (*Aequipecten*) *werthi*, echinoids, etc. which were initially referred to the Pliocene on the current evidence of the date of the well-known beds of Zanzibar; palaeontological evidence now shows that these beds also are Lower Miocene.

Other outcrops of Miocene rocks have been found inland from Dar-es-Salaam, but these are partly continental and their stratigraphy has not yet been worked out.

(c) *Mafia.*—On Mafia Island exposures of foundation rocks tend to be rare except in the north-east, being concealed by deep superficial sand over the greater part of the island and by raised reef limestones in the east. Nevertheless, rocks of Miocene date can be found at the surface in a number of places. An outcrop of richly fossiliferous marine limestone, with *Operculinella*, *Breynia carinata* and Temnopleurids indicating a Lower Miocene age, was first recognized on the shore in the south-west at Kilindoni, and found to outcrop also three miles inland at Ngombeni. Examination of seismic shot-hole samples by R. B. Holroyd then showed that clay, sandstone and limestone beds of this date are widely developed beneath the superficial deposits over most of the island. In the north-eastern peninsula an alternation of thin limestones and clays can be mapped at surface; these have yielded *Lepidocyclina* (*Eulepidina*) at two localities and are, at least in part, Aquitanian. More commonly, *Borelis melo* suggests a Burdigalian date; *Miogypsina* also occurs. Beds rich in large foraminifera occur at outcrop and at depth hereabouts; these have been identified as species of *Operculinella* and *Amphistegina*, the former of Lower Miocene age; care is needed to avoid confusion of these with small *Nummulites* in field identification.

Finally, on the western side of the peninsula at Kanga, Burdigalian limestones of different facies yield abundant *Kuphus* at high-water mark and underlie a seaward-dipping alternation of clays with thin, sandy limestones.

(d) *Kilwa.*—One Miocene outcrop was found in this area, on the islet known as Kisiwa Mbuga, 2 miles N.W. of Kilwa Masoko, where a limestone reef between tidemarks yielded the characteristic Lower Miocene *Operculinella.* The relation of this outcrop to the Eocene limestones of the Kilwa peninsula and hinterland is not apparent.

(e) *Lindi.*—The development of Miocene beds in the country between Lindi and Mikindani has long been known and a map showing the distribution with mention of characteristic fossils is included in a report by G. M. Stockley (1929). Outcrops have now been found north of Lindi, in part represented by knoll-forming limestones with *Lepidocyclina* (*Nephrolepidina*) and *Miogypsina* 10 miles north of the town and in part by a *Lepidocyclina* (*Eulepidina*) lumachelle on the north shore of the bay.

(P. E. K.)

Palaeontological Evidence in Relation to Previous Records

The detailed palaeontological examination of many hundreds of Tertiary samples from East Africa has led to the conclusion that previous conceptions of the faunal succession and of zone-fossils in the Miocene and Pliocene have to be radically revised. A number of species previously believed to characterize the Pliocene, or not to occur below the Pliocene, are now known to occur in beds of Lower Miocene age; other forms have been incorrectly identified. The reassessment of the evidence indicates, however, that the Lower Miocene and Pliocene can readily be recognized on other criteria and that some of the beds previously regarded as Pliocene are really of Lower Miocene age.

On Zanzibar, a sample near Makunduchi in the south-east was found to contain a Lower Miocene species of *Operculinella*, together with *Lepidocyclina* (*Nephrolepidina*) which confirmed the Miocene age. Later, the Lower Miocene *Lepidocyclina* (*Multilepidina*) was found near Tunguu and Dunga in the centre of the island, in association with an assemblage of fossils evidently representative of that recorded from Stockley's Locality Z.108b [1]. Furthermore, the same assemblage of fossils was found in both Stockley's cave (Locality Z.14) and above the cave on Kijini Hill. All these fossils are unquestionably *in situ*,

[1] Davies, A.M., *et alia*, (1927).

and consequently the following published records are to be regarded as of Lower Miocene, not Pliocene, age:—

LOCALITY Z.14.
Amusium n.sp.[1] (=*A.pleuronectes* (Linné) of Cox, *non* (Linné))
Chlamys (*Aequipecten*) *bornhardti* Cox
Chlamys senatoria (Gmelin)
Chlamys (*Aequipecten*) *werthi* (Philippi)
Chlamys (*Aequipecten*) *werthi* (Philippi) var. *longistriata* Cox
Spondylus insularis Cox
Carcharodon intermedius White

LOCALITY Z.108b.
Chlamys (*Aequipecten*) *pallium* (Linné)
Chlamys (*Aequipecten*) *werthi* (Philippi)
Chlamys (*Aequipecten*) *werthi* (Philippi) var. *longistriata* Cox
Clementia papyracea (Gray)
Ostrea (*Lopha*) *virleti* Deshayes (recorded as *O. hyotiformis* Philippi)
Pecten (*Oopecten*) *subflabellatus* Cox
Spondylus insularis Cox
Brissus lasti Stockley
Clypeaster sp.[2] (=*C. reticulatus* (Linné) of Stockley *non* (Linné))
Laganum sp. (= *L. depressum* Agassiz of Stockley *?non* Agassiz)
Maretia sp.[2] (= *M. ovata* (Leske) of Stockley *non* (Leske))
Metalia waylandi Stockley
Temnopleurus sp.[2] (=*T. hardwicki* (Gray) of Stockley *non* (Gray))
Temnopleurus sp.[2] (= *T. toreumaticus* (Leske) of Stockley *non* (Leske))

In the Mtoni area, just N.E. of Zanzibar City, typical Lower Miocene faunas were again found, including *Chlamys* (*Aequipecten*) *deltoidea* Cox and typical *Operculinella*, *Heterostegina* and Bryozoa, as well as a microforaminiferal fauna characterizing beds of this age. In the north, near Shangani, similar faunas are also found and it is evident

[1] All Pectinacea (*Amusium*, *Amussiopecten*, *Chlamys* and *Pecten*) are being recorded, and new forms named and described, in a paper by Eames and Cox, now in the press (*Proc. Malac. Soc., London*).

[2] Comparison, in the British Museum (Natural History), of the Zanzibar Lower Miocene *Clypeaster*, *Maretia* and *Temnopleurids* with recent material indicates that they are not attributable to the species to which they have been assigned.

that the following fauna from Locality Z.8 is of Lower Miocene, not Pliocene, age:—

Chlamys senatoria (Gmelin)
Ostrea (*Lopha*) *virleti* Deshayes (recorded as *O. hyotiformis* Philippi)
Pecten sp. (=*P. vasseli* Fuchs of Cox (*pars*) *non* Fuchs)
Paraster savignyi (Fourtau), of Stockley
Temnopleurus sp.[1] (=*T. toreumaticus* (Leske) of Stockley *non* (Leske))

In the Dar-es-Salaam district, similar foraminiferal assemblages with the Lower Miocene *Operculinella* have been found 20 miles S.E. of Dar-es-Salaam, and these fossils, together with the Lower Miocene species of *Amusium*, *Heterostegina*, *Temnopleurus*, etc., occur further south at Buyuni.

On Mafia Island, rich assemblages of *Miogypsina*, *Lepidocyclina* (*Nephrolepidina*), together with elements of the Lower Miocene faunas recorded above, have been found; the presence of *Borelis melo* (Fichtel & Moll) in a fair number of samples indicates that they belong to the Burdigalian portion of the Lower Miocene; *Lepidocyclina* (*Eulepidina*), found in only four localities, indicates, however, that some Aquitanian is present.

Near Mombasa, at Locality No. 6a[2] (cliff, N.W. of Mombasa, at Schimanzi), the same *Pecten* is found as at Locality Z.8 on Zanzibar, and it is evident that the following fauna from Locality No. 6a is to be regarded as of Lower Miocene, not Pliocene, age:—

Harpa amouretta Bolten
Antigona reticulata (Linné)
Chlamys (*Aequipecten*) *hobleyi* Cox
Chlamys (*Aequipecten*) *mombasana* Cox
Chlamys (*Aequipecten*) *nux* (Reeve)
Chlamys (*Aequipecten*) *pallium* (Linné)
Chlamys senatoria (Gmelin)
Chlamys (*Aequipecten*) *werthi* (Philippi)
Hemipecten forbesianus Adams & Reeve
Ostrea (*Lopha*) *cerata* G. B. Sowerby
Ostrea (*Lopha*) *virleti* Deshayes
Pecten sp. (=*P. vasseli* Fuchs, of Cox (*pars*) *non* Fuchs)
Plicatula exigua Cox
Plicatula mombasana Cox

[1] Comparison, in the British Museum (Natural History), of the Zanzibar Lower Miocene *Clypeaster*, *Maretia* and *Temnopleurids* with recent material indicates that they are not attributable to the species to which they have been assigned.

[2] Cox, L. R. (1930).

Similar faunas, including the *Pecten* and *Plicatula mombasana* Cox, have been found in association with the Lower Miocene *Operculinella* and other fossils further north in the neighbourhood of Kilifi Creek and Lake Bandara ya Wali. Still further north abundant *Miogypsina* occur in beds of Lower Miocene age in the Malindi area near Goshi, where beds with Pliocene fossils are also known.

The above outline will give some idea of the extent to which the Lower Miocene has been identified in the region, and of the extent to which previous ideas on the faunal succession must be modified. One result is that the " *Chlamys werthi* Beds " previously dated as Pliocene, are now known to include representatives of both the Lower Miocene and the Pliocene, the " zone-fossil " itself occurring at both levels; no evidence of beds of Middle or Upper Miocene age in the region has yet been found. There are, however, species of *Operculinella*, in association with *Pecten vasseli* Fuchs and *Ostrea* (*Lopha*) *tridacnaeformis* Cox, which are undoubtedly characteristic of the Pliocene.

(F. E. E.)

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c/o British Petroleum Co., Ltd.,
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On the Affinities of the Cretaceous Lamellibranch Genera *Trigonioides* and *Hoffetrigonia*

By L. R. Cox

ABSTRACT

The East Asian Cretaceous lamellibranch genera *Trigonioides* and *Hoffetrigonia* are discussed after examining new material of the former, and the conclusion is reached that both belong to the Unionidae and not to the Trigoniidae. A proposal to include them in a new family is withdrawn.

I. INTRODUCTION

THE genus *Trigonioides* Kobayashi and Suzuki (1936, p. 248) is known by three species, found in the Cretaceous of Korea, Manchuria, and Japan. These are (*a*) the type species, *T. kodairai* Kob. and Suz. (1936, p. 249, pl. 27, figs. 1–4 ; pl. 29, fig. 13), originally described from the Lower Cretaceous Naktong-Wakino series of Korea and N. Kyushu, Japan, and later recorded (Kobayashi and Suzuki, 1941, p. 79, figs. 1–4) from the Lower Cretaceous Talatzu series of Manchuria ; (*b*) *T. paucisulcatus* Suzuki (1940, p. 229, figs. 18–20, pl. 24), originally described as a subspecies of *T. kodairai*, from the Middle Cretaceous Siragi series of S. Korea ; (*c*) *T. matumotoi* Kob. and Suz. (1941, p. 78) (= *T. kobayashii* Matumoto, 1938, p. 14, pl. 2, fig. 2, non *T. kobayashi* Hoffet), from the Turonian Gosyonoura series of the Amakusa Is., Kyushu, Japan. Species from the Senonian of French Indo-China included by Hoffet (1937) in *Trigonioides* were referred by Suzuki (1940, p. 229) to a new genus, *Hoffetrigonia*. *Trigonioides* and *Hoffetrigonia* were included by their authors in the Trigoniidae, although they occur in fresh- or brackish-water beds. In my review of that family (Cox, 1952), however, I considered, from the published information about their hinge-structure, that they should be " united to form a further family, the Trigonioididae, which seems to be more appropriately placed near the Mutelidae in the Unionacea than in the Trigoniacea." On the other hand, Dr. F. Haas, of the Chicago Natural History Museum, a leading authority on the Unionacea, told me (*in litt.*) that, in his opinion, they " seem to be Trigoniidae, not quite typical ones, perhaps, but certainly more typical for Trigoniidae than for Unionidae." Professor T. Matsumoto kindly brought to England and allowed me to examine specimens of the type species of *Trigonioides* (*T. kodairai*), collected by Y. Ohta from one of its original localities, situated near Rikimaru, in the Fukuoka Prefecture of Kyushu, and belonging to the Geological Institute of Kyushu Imperial University. He informed me that the type specimens of the species were destroyed during the last war.

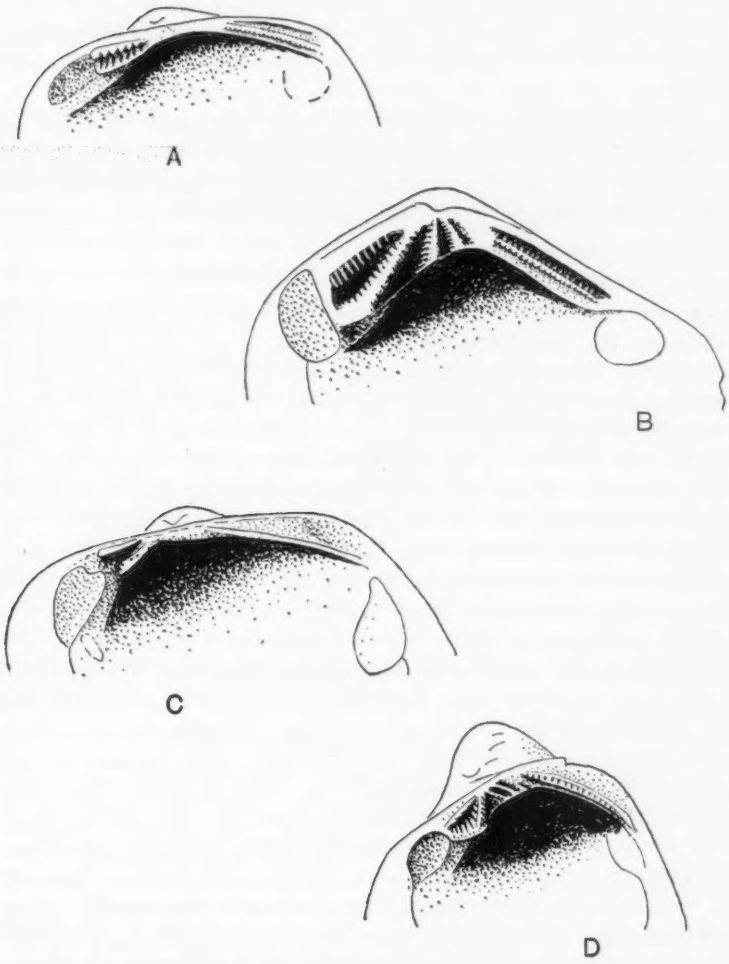
II. Characters of Trigonioides

No good figures of the hinge-structure of the three known species of *Trigonioides* have yet been published, and existing descriptions are not very clear, probably owing to the imperfect nature of the material. Kobayashi and Suzuki (1936, p. 248) originally wrote, "The hinge teeth are composed of two laterals divergent from the beak in a wide V-shape and probably bifid cardinals inserted between, among which the anterior cardinal is smaller than the posterior." They commented upon the strong crenulation of the teeth. Their subsequent description (1941, p. 78) of specimens of the type-species from Manchuria did not add to our knowledge of the dentition. Matumoto (1938, p. 14) wrote of the Turonian species later named *T. matumotoi*, "Right valve provided with two strong teeth crossed laterally by numerous small ridges, among which the anterior is inclined obliquely forward and downward from the top of the beak. Left valve with three teeth, among which the median is the strongest and crenulated, the anterior not small and also crenulated, and the posterior is very small." Suzuki (1940, p. 230) described the dentition of *T. paucisulcatus* as follows: "Hinge teeth three on each valve; two anterior cardinals subparallel to each other, feebly curving, fairly long and solid, distinctly and acutely crenulated; one posterior cardinal widely apart from the anterior ones, nearly straight, a little longer and far more weakly crenulated than the anterior."

The specimens of *T. kodairai* which I have examined are preserved, partly as internal and external moulds of the shell, in a hard slaty rock. They have the characteristic surface ornament of the species and there can be no doubt that they are correctly identified. The impressions of the hinge-teeth are clear, particularly in the case of two right valves, and artificial casts well reproduce the original teeth. The dentition and musculature of the shell may be described as follows.

Right valve (Text-fig. 1A) with three teeth, consisting of (*a*) a narrow, elongate tooth with regular transverse crenulations on both sides, parallel with and extending the whole length of the postero-dorsal margin, (*b*, *c*) two moderately elongate but more prominent teeth originating a short distance anterior to the umbo, remaining subparallel with the antero-dorsal margin, and enclosing a deep recess with transversely crenulated walls for the reception of the corresponding tooth of the other valve. There are no distinct subumbonal teeeth, but immediately below the umbo is a shallow triangular recess, the nature of which is doubtful.

Left valve with the following three teeth: (*a*, *b*) two narrow, elongate posterior teeth, of which the upper is crenulated on both sides and extends along the whole length of the margin, and the lower is crenulated on its upper side only and extends only half-way to the umbo



TEXT-FIG. 1.—Hinge-structure (× 1) of right valves of the following:—

A. *Trigonioides kodairai* Kob. and Suz., Lower Cretaceous, N. Kyushu, Japan (based on an artificial cast made from a natural mould).

B. *Hoffetrigonia trigona* (Hoffet), Senonian, Bas-Laos (based on a text-figure and photograph published by Hoffet).

C. *Unio* (*Lampsilis*) *cariosus* Say, Recent, N. America (three teeth are present, as in *Trigonioides*).

D. *Castalia ambigua* Lamarck, Recent, S. America (subumbonal teeth are present, as in *Hoffetrigonia*, in addition to the three teeth found in *Trigonioides* and *Unio cariosus*; the teeth are strongly crenulated, as in *Trigonioides* and *Hoffetrigonia*).

from the posterior end of the margin, (*c*) a prominent, sharp-edged anterior tooth, crenulated on both sides, which is subparallel with the antero-dorsal margin and fits into the corresponding recess in the other valve.

The deep anterior muscle scar (well seen in the artificial cast of the right valve) is placed very dorsally, immediately in front of the anterior teeth, and on its posterior side is a slight thickening of the wall of the shell, extending towards the umbo below these teeth.

III. Affinities of Trigonioides

Comparison of its hinge-structure with that of a typical Unionid (Text-fig. 1C), can, in my opinion, leave no doubt as to the Unionid affinities of *Trigonioides*. The arrangement of the hinge-teeth is just the same, the two anterior teeth of the right valve being identifiable (in the Bernard notation) as 3*a* and 5*a*, and the posterior tooth as 3*b* ; while in the left valve the strong anterior tooth is 2*a* (4*a* being obsolete or almost so), and the two posterior teeth 2*b* and 4*b* (compare H. Douvillé, 1913, p. 442, fig. 17 ; A. M. Davies, 1935, p. 132, fig. 175). The Trigoniid hinge, on the other hand is characterized by the great reduction of the upper anterior tooth (5*a*) in the right valve, and by the fusion in the left valve of the two teeth 2*a* and 2*b* to form a single bifid tooth, 2.[1] It was, no doubt, the very strong transverse crenulation of the teeth in *Trigonioides* and the imperfect condition of the original material on which the genus was founded which led its authors to refer it to the Trigoniidae. Many species of the Unionacea, however, have transversely crenulated teeth, this character being particularly well marked in the genus *Castalia* Lamarck (Text-fig. 1D).[2] The steep, oblique ribs, which meet along the middle of the valve to form acute V's, are another feature of *Trigonioides* which suggested comparison with some groups of the Trigoniidae. Similar ornament may, however, be present in the living Unionid genus *Prisodon* Schumacher (*Hyria* Lamarck), and also occurs in *Nippononaia* Suzuki (1941, p. 411), from the Lower Cretaceous of Japan, regarded by its author as a subgenus of *Unio*. It is now clear that *Trigonioides* is quite a normal representative of the Unionidae and that my proposal to include it in a distinct family Trigonioididae within the Unionacea must be withdrawn. *Nippononaia* differs from it in that its steep surface ribs extend to the posterior margin instead of being replaced posteriorly by broader and

[1] See Cox, 1952, p. 47, top two figures. (By an error in the lettering, this tooth was indicated as 2*b*.)

[2] Simpson (1914) includes *Castalia*, *Hyria* (= *Prisodon*), and other genera of the subfamily Hyriinae in the Unionidae, but Thiele (1934, p. 838) includes this subfamily in the Mutelidae.

more nearly horizontal ribs, in the much finer crenulation of its anterior teeth, and in the smoothness of its posterior teeth.

IV. Characters and Affinities of Hoffetrigonia

The genus *Hoffetrigonia* was founded by Suzuki (1940, p. 229) on certain species from Senonian brackish-water beds in French Indo-China which Hoffet (1937) had referred to *Trigonioides*. The hinge-structure of Hoffet's specimens (said to be at Hanoi) was much better preserved than that of the original specimens of *Trigonioides*, and his figures (on which text-fig 1B is based) are fairly clear. As in that genus, there are strong, rather elongate, anteriorly sloping teeth, which may be designated as 3*a* and 5*a* in the right valve and 2*a* in the left, and elongate, crenulated posterior teeth, 3*b* in the right valve and 2*b* and 4*b* in the left. In addition, however, there are narrow, steep subumbonal teeth, two in each valve, for which Bernard's notation has no appropriate designations. Comparable subumbonal teeth are present in many Unionidae (for example, *Castalia*, text-fig. 1D) in which elongate posterior and anterior teeth are also found. Although often weak, they become as prominent as in *Hoffetrigonia* in some genera (for example, *Quadrula*). It is important to note that the stout median tooth (2) which is characteristic of the left valve of all genera of the Trigoniidae is absent in *Hoffetrigonia*. The hinge-structure of this genus thus also appears to be fundamentally Unionid and to have no close affinities with the Trigoniid hinge.

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Department of Geology,
British Museum (Natural History).

CORRESPONDENCE

UPPER YOREDALES AND MILLSTONE GRIT RELATIONS IN THE STAINMORE COALFIELD

SIR,—In his interesting account of the Stainmore Coalfield (*Geol. Mag.*, xcii, 1955, 218–230), Dr. T. D. Ford claims that the Millstone Grit there oversteps the Mirk Fell Beds and "Fell Top" (*recte* Stonesdale) Limestones of the outlier, and that the limestone-outcrops at the mouth of Craghouse Gill, referred by me (*Proc. Geol. Assoc.*, xlvi, 1935, 124) to the "Fell Top" Limestones, lie above the lowest member of the Millstone Grit.

There are good sections of the Upper Yoredales and early Millstone Grit a short distance south-east of the Stainmore outlier, both north and west of Nine Standards. This ground has been revised of recent years by Dr. A. J. Rowell and myself and it is hoped to publish an account of our work shortly. One result of it has been the recognition of the importance of a marine horizon between the Upper Little and Crow Limestones for which we propose the name Faraday House Marine Band : it overlies Tiddeman's (*Mallerstang Memoir*, 1891, 83) "lower flags" above the Upper Little Limestone and was described by him as having "a brown gingerbread-like top with fossils". This crosses Faraday Gill under Nine Standards, close to a ruined hut known as Faraday House. My failure to recognize the importance of this persistent marine horizon, which can be followed far to east and south, led to serious errors along the southern edge of my Stainmore map.

In the Stainmore outlier, Dr. Rowell has recognized the Band in Coldkeld Beck and Intake Gill in the south, and in Mousegill Beck due south of Stricegill Farm, in each case a little below the sandstone taken by Dr. Ford as the basal Millstone Grit. This sandstone is, however, the local representative of the Ten Fathom Grit for which we propose the name Uldale Sill, from Ul Dale, just over the Yorkshire boundary at the head of Birkdale, since it corresponds to the upper part only of the Ten Fathom Grit of Swaledale, the Faraday House Marine Band lying in the middle of the latter.

Above the Uldale Sill follows the Crow Limestone, both in the outlet and on the main outcrop. This is succeeded by a thick mass of shales in which lie (in ascending order) the Lower Stonesdale Limestone, the Upper Stonesdale Limestone, and the Mirk Fell Gannister. It is these limestones which outcrop at the mouth of Craghouse Gill, their section agreeing well with that seen in Ul Dale :

Craghouse Gill.	*Ul Dale.*
Shales (cut off by fault).	Shales, 30 ft.
Limestone, 1 ft 6 in.	Limestone, 2 ft.
Sandstone, 3 ft.	
Shales, 18 ft.	Shales, 24 ft.
Limestone, sandy below, 1 ft. 10 in.	Limestone, 2 ft.
Shales (cut off by fault).	Shales, 14 ft. 6 in.
	Crow Limestone.

Contrariwise, such limestones are never seen between the first and second lowest members of the Millstone Grit of the Askrigg Block. In brief, Dr. Ford has misidentified the Uldale Sill with the basal Grit and the great overstep claimed by him does not exist.

I note that Dr. Ford applies the name Dent Fault to what I have termed (op. cit., 132–7) the Boundary Fault of the Coalfield. The true Dent Fault disappears under the New Red beds at Thringill, south of Kirkby Stephen, and there is no justification for identifying it with a fault several miles away and with which it has no physical connection.

J. SELWYN TURNER.

DEPARTMENT OF GEOLOGY,
UNIVERSITY OF LEEDS.
9th July, 1955.

REVIEWS

GRAPHIC PROBLEMS IN PETROLEUM GEOLOGY. By L. W. LE ROY and J. W. LOW. pp. 238, with 43 figs. Harper and Brothers, New York. Hamish Hamilton, Ltd., London, 1954. Price 36*s*.

Thirty-one problems with a great deal of introduction, explanation, and illustration are arranged so as to instruct by practice. The loose-leaf plastic spine and quarto size of page allows the book to open flat for drawing or abstracting data from tabulated well logs. Most problems would take several hours to complete properly, for even though the essence may be grasped in a few minutes the whole emphasis is on careful drawing of detail to produce finished work. Many of the problems involve different graphical treatment of scores of well logs from systematically drilled fields and it is only in problem 20 that a structural contour map is obtained from topography and outcrop. Here then is a useful counterbalance to the usual university mapping course in this country. There is much emphasis at the beginning on drafting various columns and sections with standard ornament and colours followed by the construction of simple " fence " (map-panel and isometric-panel) diagrams. Material is then provided for plotting a variety of lithofacies maps and sections from isopach data, lithological analyses, etc. The remaining third of the book gives exercises in the application of simple structures to petroleum geology, including " Busking " folds, volumetric calculations from contour maps, and brief treatments of stereographic projections and air photographs.

This book should serve its purpose but will be of little direct use to academic students because of its price, the time taken to do any problem, and the very small intellectual content of the many pages of platitudinous and often pompous exposition. It will, however, give an insight into one aspect of industrial " know-how " and provide examples which have been carefully prepared by experienced petroleum geologists. It is also a useful reminder that a large concern requires from its geologists considerable experience and discipline in drawing as well as a research outlook.

W. B. H.

INTERNATIONAL SEDIMENTARY PETROGRAPHICAL SERIES. Vol. I—FAZIES UND MIKROFAUNA DER GESTEINE DER BAYERISCHEN ALPEN. By HERBERT HAGN. 8vo, pp. 11 + 174; with 141 micrographs in 71 pl.; 2 text-figs.; 8 stratigraphic tables. E. J. Brill, Leiden, 1955.

An analysis of the literature of sedimentology reveals increasing interest in the subject, with particular emphasis upon the relationships between petrography and depositional environments.

More of the effort has been concentrated upon unconsolidated sediments which lend themselves to quantitative methods of analysis, than the harder rocks, including the important carbonate group, yet these reveal in sections under the microscope a much wider range of " microfacies " [1] variations, eloquent of depositional and post-depositional conditions, than is to be found among the soft clastics, while they may also contain microfossils giving direct evidence of age.

This fact has long been accepted and applied by Petroleum Geologists, working in hard rock areas, who have recognized the significance of microfacies in the stratigraphic approach to oil-finding and their value as criteria for bed-by-bed correlation between outcrops and wells, by methods which are applicable alike to hand-specimens and to drill cuttings.

Unfortunately, detailed records of microfacies are complex, qualitative, and not easy to communicate except by illustration in micrographs. This

[1] CUVILLIER, J., and V. SACAL. Corrélations Stratigraphiques par Microfaciès en Aquitaine Occidentale. E. J. Brill, Leiden, 1951.

method is followed successfully by Cuvillier in his book on the microfacies of Aquitaine (op. cit.), which attracted wide interest at the Third World Petroleum Congress (1951) where it was presented; and from it the idea was born of a bilingual " International Sedimentary Petrographical Series " of companion volumes representing geological provinces in other parts of the world.

This venture will be welcomed by all petrographers and stratigraphers, because it promises to give proper attention to the neglected harder rocks in the realm of sedimentology; and because it will facilitate instructive world-wide comparisons of sedimentary rock-types of equivalent ages and facies. For the general geological reader it will provide also, handy synopses of regional stratigraphy.

Dr. Herbert Hagn's volume I, on the sedimentary microfacies of the Bavarian North Alps, gives an excellent start to the new series.

It contains 141 high quality half-tone micro-photographs of characteristic rocks, ranging in age from Triassic to Miocene, and accompanied by adequate notes in German and English on their sources, diagnostic features, and stratigraphic contexts.

The stratigraphical succession of the region is conveniently summarized in eight tables correlating world-wide time-stratigraphic divisions with provincial rock-stratigraphic units which are referred to by their local terms.

Descriptions of these lithological units are dispersed among the remarks on relevant micrographs and could with advantage have been brought into closer connection with the tables by repetition in them, or by suitable cross-referencing.

Regional perspectives are maintained in a short introductory account (in German only) of the tectonic-stratigraphic evolution of the Bavarian Alps. This is illustrated by a small-scale map, and by a series of diagrammatic profiles showing the successive development of major basins and swells, which divided the region into tectonic-stratigraphic belts.

The arrangement and printing of the volume are very good; and the English translations are adequate, though certain terms which may be unfamiliar to some English readers (Bankkalk; Knollenkalk; etc.) might have been substituted or explained.

F. R. S. H.

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